

**FLASH FLOODING IN EASTERN KENTUCKY:  
AN ANALYSIS OF THE 3-4 AUGUST 2001 EVENT**

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**FLASH FLOODING IN EASTERN KENTUCKY:  
AN ANALYSIS OF 3-4 AUGUST 2001**

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# FLASH FLOODING IN EASTERN KENTUCKY: AN ANALYSIS OF THE 3-4 AUGUST 2001 EVENT

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Limited research exists concerning flash flooding in the United States. However, flash floods can occur anywhere and cause more fatalities than any other weather related natural disaster. Eastern Kentucky experienced an average of 41 flash floods per year (1993-2002), yet little research exists for this region. Therefore, this paper presents an analysis of the particularly devastating flash flood event of 3-4 August 2001, which resulted in \$15 million worth of property damage and two deaths.

Previous studies indicate that flash floods typically occur under relatively 'benign' conditions, the most common characteristics being a quasi-stationary system and high atmospheric moisture. The slow moving nature of these systems, coupled with low dewpoint depressions, leads to the overabundance of precipitation that produces flash floods.

Unfortunately, this relatively 'benign' nature of weather systems poses a challenge for accurate forecasting of flash flooding. Furthermore, severe weather threats, such as tornadoes, tend to overshadow the threats posed by flash flooding, sometimes interfering with the timely issuance of flash flood warnings.

To improve the accuracy of flash flood forecasting, several methods have been developed. For instance, lightning strike density helps forecasters in the West to locate regions more likely to receive the heavy rains necessary for flash flooding. Numerical forecasting models also help forecasters, especially when necessary data does not exist for the forecast area. One

such model is the Penn State/ NCAR Mesoscale Model MM5. This model consists of several options allowing for variations in climate and terrain. MM5 was used to simulate the 3-4 August 2001 flash flood event, since upper air data does not exist for Eastern Kentucky. The model used for this paper utilized the Grell parameterization scheme, which provided better results than the other parameterization schemes tested (Kain-Fritsch).

The Eastern Kentucky flash flood event, like the majority of other flash flood events throughout the Contiguous United States, did in fact occur under relatively benign conditions. This event occurred as heavy rains fell ahead of a trough axis associated with a slow-moving cold front. As the west-east oriented front moved southward, it stalled along the Appalachians and pivoted, becoming southwest-northeast oriented. High atmospheric moisture existed from the surface through 500-mb ahead of the front. The winds were weak and from a northerly direction, while the stability indices indicated marginal instability throughout the event. About half of the flash floods reported during this event occurred prior to the issuance of a flash flood warning, while no record of a flash flood watch existed for any of the counties prior to flash flooding. Accurate forecasting of flash floods during this relatively benign event was further hindered by the lack of a flash flood climatology for the region.

## I. INTRODUCTION

Flash floods cause more fatalities annually than any other weather related natural disaster (Maddox *et al.* 1979; Doswell *et al.* 1996; AMS 2000). A flash flood is “a sudden and destructive FLOOD occurring with little or no prior warning caused usually by excessive rainfall and/or rapid snow and ice melt” (Smith 2001, 86). Flash floods occur quickly, usually in less than six hours, and typically affect an area no larger than 1,000 km<sup>2</sup> (AMS 2000). Flash floods most commonly occur as a quasi-stationary system moves slowly through an area. Generally, the training of several embedded thunderstorms lead to the excessive rainfall necessary for flash flooding (Doswell 2003).

Hydrological and meteorological conditions combine to produce flash flooding. Hydrological factors, including slope, vegetation, human activities, and soil conditions, determine how rainfall will behave once on the ground (Doswell 2003). This research, however, focuses on the *meteorological* factors, particularly atmospheric moisture, which lead to the excessive rainfall resulting in flash floods.

Much of the research previously conducted indicates that flash flooding generally occurs during relatively benign conditions, which make flash flood forecasting difficult (Maddox & Grice 1986; Doswell *et al.* 1996). Although several studies exist concerning flash floods throughout the United States, little research is conducted regarding flash flooding in the southern Appalachians. Therefore, this research examines the meteorological conditions of an eastern Kentucky flash flood event to determine if it also occurred under similarly benign conditions.

The first part of this thesis highlights research conducted by others concerning several of the major flash flood events, which have occurred in the Contiguous United States (Chapter 2). Through examination of numerous regional flash floods, researchers have built flash flood

climatologies, identifying some of the challenges involved in flash flood forecasting, as well as some of the methods developed to improve forecasting (Chapter 3). For instance, the use of mesoscale models, such as MM5, result in more accurate flash flood forecasting (Chapter 4).

The second part of this thesis focuses on flash flooding in eastern Kentucky, with an emphasis on the flash flood event of 3-4 August 2001. Although national atmospheric data is available from many different sources, available data for eastern Kentucky is quite limited. Consequently, the MM5 model provided simulations of the atmospheric conditions during 3-4 August 2001 (Chapter 5). In addition, simple flash flood and precipitation climatologies were created from existing data for 1990-2002, which provide a general picture of flash flooding in eastern Kentucky (Chapter 6). A detailed examination of the August 2001 flash flood event was conducted at the large-scale and the local-scale (Chapter 7). Finally, this research concludes with an analysis of how this eastern Kentucky flash flood event compares to the other major events examined in this paper (Chapter 8).

## II. SELECTED FLASH FLOOD EVENTS AND KEY FEATURES

Examining previous flash flood events provides information regarding atmospheric features related to flash flooding. This chapter examines various evolving storm systems that have occurred in the contiguous United States, resulting in heavy rains and subsequent flash floods. An understanding of these previously analyzed systems helps to identify key characteristics that might also be associated with events found in eastern Kentucky.

### 2.1 South Dakota

South Dakota experienced its worst flash flooding in recorded history on 9 June 1972, when orographic forcing of very moist air produced heavy precipitation along the eastern edge of the Black Hills (Maddox *et al.* 1978; Nair *et al.* 1997). The flooding resulted when a polar front pulled extremely moist air into the area from the northeast. This storm contained two frontal structures associated with a polar air mass (Maddox *et al.* 1978), and resembled a tropical system rather than a midlatitude system (Nair *et al.* 1997).

At 12Z 9 June, a negatively tilted ridge existed at 500-mb, with a short-wave trough to the south-southwest of the area. The sounding for Rapid City SD indicated a shallow layer of moist air below a strong temperature inversion at 860-mb. The lifted index indicated stable conditions with a value of +6. The sounding for Huron indicated very moist conditions up to 780-mb, with a precipitable water value of 33.5 mm (1.32 in) and an unstable atmosphere with a lifted index of -7 (Maddox *et al.* 1978).

As the unstable air moved over the Black Hills around 00Z 10 June, orographic lifting broke the strong temperature inversion that existed, resulting in very heavy rains (Maddox *et al.* 1978; Nair *et al.* 1997). The Rapid City sounding indicated moist, unstable air with a precipitable



water value of 19.3 mm (0.76 in) and a lifted index of  $-5$ . Within a four-hour period, the Rapid Creek drainage basin received a maximum of up to 381 mm (15 in) of precipitation, which led to the devastating flash floods (Maddox *et al.* 1978).

## 2.2 Colorado

On 31 July 1976, devastating flash flooding resulted when a quasi-stationary thunderstorm complex moved over the Big Thompson drainage basin (Caracena *et al.* 1979; Maddox *et al.* 1978). The thunderstorm complex consisted of two frontal structures; a leading front and a trailing front. Shifts of wind accompanied the leading front, while a strong temperature gradient and stronger winds accompanied the trailing front (Maddox *et al.* 1978).

At 12Z on 31 July, moist conditions existed over the eastern slopes of the Rockies and a negatively tilted ridge dominated the upper air (Maddox *et al.* 1978). A 500-mb short-wave trough existed on the western side of this ridge (Caracena *et al.* 1979). The Denver sounding indicated a very moist lower and middle troposphere, capped at 670-mb. The precipitable water value was 50 mm (1 in), a value 50% above normal for July. The Sterling CO sounding also indicated above normal precipitable water values. At 1920Z, Sterling represented the atmosphere behind the trailing front, indicating conditionally unstable air and a very high moisture content, with a 33.3 mm (1.31 in) precipitable water value (Maddox *et al.* 1978).

By 00Z 1 August, the trailing front merged with the leading front, strengthening the system (Maddox *et al.* 1978). As the trailing front moved westward, the moist air behind the front moved upslope, resulting in very heavy rains of more than 304.8 mm (12 in) over the Big Thompson drainage basin (Maddox *et al.* 1978; Caracena *et al.* 1979). The National Weather Service recorded a 48-hr total of approximately 280-300 mm (11-12 in) of rain through a bucket survey near Glen Comfort and Glen Haven, a total five to six times greater than the monthly average for the area (Caracena *et al.* 1979).

On 28 July 1997, Colorado experienced another major flash flood as a sequence of rainstorms moved through Fort Collins, accompanying a cold front that moved in from the northeast. Colorado State University, in Fort Collins, recorded 134.6 mm (5.3 in) of rain in less than six hours, exceeding both the 100-yr 6-hr (89 mm; 3.5 in) and the 100-yr 24-hr (122 mm; 4.8 in) totals. The majority of the precipitation fell to the west and northwest of Fort Collins, with a total of 259 mm (10.2 in) recorded just southwest of Fort Collins. This event resulted in “one of the largest rainfall events ever documented over a developed urban area in Colorado” (Petersen *et al.* 1999, 191).

Rainfall began by the evening of the 27<sup>th</sup>, with 61.5 mm (2.4 in) reported west of Fort Collins. During the morning of the 28<sup>th</sup>, weak upper-level winds and moist air flowing up-slope set the stage for heavy rains and flash flooding. Northwest of Fort Collins, Laporte CO reported an accumulation of 177.8-228.6 mm (7-9 in) of rain as the rain ended around noon on the 28<sup>th</sup> (Petersen *et al.* 1999).

Another period of rain began around 1700 MDT, again over the area just west - northwest of Fort Collins. Heavy rains fell over portions of Fort Collins as two small convective systems moved over the area. These systems were followed by a third, as the winds became more easterly, increasing the moisture availability over Fort Collins. As the system moved in from the southwest, it became quasi-stationary over western Fort Collins. The heavy rains reached 100-125 mm (4-5 in) of rain per hour, falling on already saturated ground. The resulting runoff and overflowing rivers and streams led to the closing of several roads. The rain ended around 2300 MDT, leaving behind flooded homes, streets, and a derailed train (Petersen *et al.* 1999).

The Denver sounding for 18 MDT indicated a moist atmosphere. The precipitable water value was 34 mm (1.3 in), which doubled the average of 17 mm (.67 in) for July. The dewpoint depressions were small and the lifting condensation level was low at 764-mb. The lifted index of -2.8 indicated a moderately unstable atmosphere. Weak southwesterly winds dominated the mid-troposphere (Petersen *et al.* 1999).

The synoptic features of this flash flood event were very similar to those of the Big Thompson Storm in July 1976 and the Rapid City storm in 1972. However, the lifted index indicated a more stable environment than occurred during the Big Thompson and Rapid City storms (Petersen *et al.* 1999). The devastation experienced by this event led to an improvement of the Fort Collins' emergency warning system, which decreased the amount of damage caused by their next flash flood event, in 1999 (Weaver *et al.* 2000).

### 2.3 Virginia

On 27 June 1995, Madison County, Virginia experienced devastating flash flooding when a weak cold front moved into the area from the north. Significant damage occurred when an average of 334 mm (13 in) of rain fell on already saturated ground, leading to mudflows and flash flooding (Pontrelli *et al.* 1999).

During the day prior to the flash floods, a large-scale trough existed over the Mississippi River Valley and a ridge axis existed over the Northeast. The Appalachians caused divergence over the Ohio Valley and several short-wave troughs moved along the long-wave trough. A surface high over the New England area brought cold air southward (Pontrelli *et al.* 1999).

The lifted index was negative, at approximately  $-1$ . The CAPE was weak at 150 J/kg and the totals index predicted possible thunderstorms. The winds were light at all levels and the atmosphere was very moist. The precipitable water values were calculated at about 50 mm, 160% above normal. The surface dewpoint temperature was recorded at 23 °C and the 850-mb dewpoint temperature was recorded at 14 °C (Pontrelli *et al.* 1999).

A long-wave trough and three short-wave troughs characterized the atmospheric conditions during the day of the flash flood. Cold air damming occurred because of increasing stability in the New York and New Jersey areas. Northern Virginia, on the other hand, maintained moist conditions with precipitable water values close to 50 mm. The high

precipitable water and the low CAPE, along with weak mid-level steering winds, contributed to the slow-moving system leading to heavy rainfall (Pontrelli *et al.* 1999).

Two systems resulted from different air masses with different thermodynamic properties: the Piedmont and the Madison systems. Because of the cloud cover produced from the anvil outflows of the two systems, the surface temperature rose little from the overnight low. As the cool, moist air moved southward, it encountered a warmer air mass where the temperatures quickly climbed into the 30's °C by early afternoon. However, the Sterling station failed to represent the clearer skies and warmer temperatures of southeastern Virginia (Pontrelli *et al.* 1999).

The Madison system propagated toward the southwest at 1.2 m/s, while the Piedmont system propagated more rapidly toward the southeast at 3.3 m/s. The atmosphere was extremely moist, with dewpoint depressions of only a few degrees characterizing the inflowing air. The slower propagation speed of the Madison system resulted in precipitation maxima that were two to three times greater than those reported for the Piedmont system (Pontrelli *et al.* 1999).

By 00Z 28 June, the stability at Sterling had increased with a CAPE of 30 J/kg and a positive lifted index value. The axis of the third short-wave trough passed over Madison County at 21Z. At approximately the same time, the Madison system and the Piedmont system both dissipated (Pontrelli *et al.* 1999).

The Madison system exhibited similar characteristics to the Rapid City SD storm, the Big Thompson Storm in Colorado, and the Fort Collins CO flash floods. In each case, a negatively tilted trough-ridge pattern straddled a mountain range and a weak short-wave trough approached from the southwest. A slow-moving cold front and moist easterly winds flowing perpendicular to the mountains characterized each event. Weak downdrafts existed for these events, allowing the systems to move slowly and produce heavier precipitation amounts (Pontrelli *et al.* 1999).

## 2.4 Texas

Forecasters in south Texas are often challenged by heavy rains and flash flooding. In fact, about 20% of the national flash flood total occurred in south Texas, with the south Texas Forecast Office in San Antonio reporting more flash floods than any other office in Texas. Weak tropical storms usually cause heavy rains and flash flooding during the warm months, with rainfall amounts usually exceeding 500 mm. Of course, flash flooding can also occur without the presence of a tropical storm (Read and Maddox 1983).

Austin lies just to the east of an area known as the Texas Hill country, in Central Texas. The change in elevation and the proximity to the Gulf of Mexico make Austin especially prone to heavy rainfall and flash flooding (Maddox & Grice 1986). In May of 1981, Austin experienced a particularly devastating flash flooding.

On 24 May 1981, the large-scale meteorological conditions were normal for the spring season over the southern Plains. The previous three nights had experienced convective storms as a cut-off low slowly moved across the northern Plains. The winds were weak at the surface and 500-mb level, and the height gradients changed slowly. At 850-mb, a moisture tongue accompanied the influx of warm air into central Texas. The temperatures increased towards the southwest and the subtropical jet rested over southern Texas. At the 200-mb level, diffluent flow existed preceding a short-wave trough (Maddox & Grice 1986).

All the large-scale conditions were common for the time of year. No major fronts existed in the area, and the atmospheric conditions present at the large-scale provided few useful clues with which to explain the cause of the flash flooding. The mesoscale conditions, however, turned out to be much more complex. Satellite images revealed thunderstorm activity scattered throughout the south-central United States prior to the flash flooding in Austin. A mesoscale convective system (MCS) moved into the moist, unstable atmosphere above Texas leaving a brief period of stability. However, by afternoon, the atmosphere became unstable again, this time providing the conditions necessary for flash flooding. Nevertheless, the forecasters had

determined the area out of danger for flooding and focused instead on the severe weather (intense thunderstorms, winds, and tornado threats) that accompanied the storms (Maddox & Grice 1986).

Austin received between 75-125 mm (3-5 in) of precipitation with this event, saturating the ground in the Hill Country. Because of the rainfall received there over the previous nights, forecasters were aware of the potential for flooding and had examined the flood potential of the area, in case more rain fell. However, because of the severe weather distraction, the flash flood warning for Austin did not come until after the heavy rains were near their end to the west, and flash flood reports had already begun (Maddox & Grice 1986).

Another devastating flash flood occurred in Texas on 10-14 June 1981, when 125-250 mm (5-10 in) of precipitation caused flooding of all the major river systems in Texas. One location even reported over 500 mm (20 in) of precipitation for this event (Read and Maddox 1983). On 9 June 1981, a dry and stable atmosphere existed over Texas. According to satellite imagery for 2030Z, no deep convection existed over the southern portion of the state. In fact, due to a shallow marine layer, few fair weather clouds even existed. The 00Z 10 June soundings indicated that the lower troposphere was very stable and the K-index was below 10 throughout south Texas (Read and Maddox 1983).

However, by 12Z, sounding data indicated that the surface to 700-mb moisture had dramatically increased, and the K-index had risen to the mid-30's in the south Texas region. At 850-mb, a 10-15 m/s southerly wind blew moisture from the Gulf of Mexico over south Texas, and there existed a weak horizontal wind shear, with no obvious warm advection. A weak trough with poorly defined circulation features existed over south Texas at the 500-mb level. Analysis of the 200-mb level revealed a zonal flowing polar jet far north of Texas and a weak subtropical jet south of Texas. Diffluence was visible over the eastern portion of Texas, and only the upper Texas coast experienced isolated showers (Read and Maddox 1983).

However, deep convection rapidly developed during the day of 10 June. In the afternoon, a large cluster of thunderstorms developed with a well-defined outflow boundary, along which

new storms developed during the late afternoon and into the night. These storms resulted in rainfall amounts between 125-175 mm (5-7 in) over the coastal plains region. The area around Austin reported precipitation amounts up to 175 mm (Read and Maddox 1983).

By 12Z 11 June, a stronger moist southerly flow existed at the 850-mb level and a well-defined, closed low existed at the 500-mb level. At 200-mb, a cyclonic circulation, which had remained nearly stationary, had intensified over south Texas. Meanwhile, an anticyclonic circulation developed over southern Mississippi, increasing the wind speeds over eastern Texas. The changes in wind speed were similar to those associated with a tropical disturbance over the Gulf of Mexico. Strong diffluence existed where thunderstorms occurred. Visible on the infrared satellite images, leftover cirrus existed above eastern Texas and the northwestern portion of the Gulf of Mexico, while convection continued southeast of the cirrus debris. The atmospheric patterns visible at 12Z 11 June continued with little change until 14 June, producing thunderstorms and rain near the center of the mid-tropospheric low and in the zone of upper-level diffluence (Read and Maddox 1983).

## **2.5 Missouri**

On 12 September 1977, Kansas City MO experienced two extreme precipitation events resulting in major flash flooding. Although many of the major streams in the area flooded, the most devastating flooding occurred in the Brush Creek Basin. This event resulted in record high water with a peak flow of  $17,600 \text{ ft}^3 \text{ s}^{-1}$  (Hales 1978).

The first heavy rainstorm occurred during the morning hours of the 12<sup>th</sup>, producing 180 mm of precipitation over dry ground during a six-hour time span. No major flooding accompanied this storm. However, fifteen hours later, another rainstorm moved through the area producing around 200 mm of precipitation, with approximately 140 mm of rain accumulating within two hours. The heavy rains increased the Brush Creek flood crest by three times what it had been that morning (Hales 1978).

At 1900 CDT on the 12<sup>th</sup>, a 500-mb short wave trough existed over the western Kansas border. Over the eastern Kansas border, an area of positive vorticity advection existed, and synoptic-scale convergence suggested the existence of frontogenesis. At 850-mb, an area of strong convergence existed over the northern Kansas-Missouri border with mixing ratios greater than 16 g/kg<sup>3</sup>. The area of warm air advection had shifted from its location twelve hours earlier over northeast Kansas into northwest Missouri (Hales 1978).

Topeka had a lifted index of -7, indicating very unstable air. In fact, thunderstorms were present just northwest of Topeka. At the surface, strong convergence existed where widespread thunderstorm activity occurred. These thunderstorms created strong temperature and wind boundaries, which moved toward the north and were clearly indicated in satellite imagery (Hales 1978).

At the mesoscale, a surface low moved to the east-northeast, creating a nearly stationary boundary where the warm front met the outflow boundary from earlier storms. Along this boundary, new thunderstorms quickly developed, and by 2000 CDT, heavy rain began to fall over the Brush Creek drainage basin. Southwest of Kansas City, new storms continued to develop, while other storms developed along a cool air outflow boundary and moved slowly to the south. By 2100 CDT, the storms began traveling towards the north, where they merged with the already northward moving storms. From 2000 to 2100 CDT, the wind southwest of Topeka increased to more than double their original speed, though the temperature and dewpoint changed only slightly. Nevertheless, the resultant thunderstorms most likely occurred due to an increase in convergence (Hales 1978).

The new line of thunderstorms was visible on the infrared satellite imagery. Thunderstorms associated with the front, however, remained west of Topeka. By 2200 CDT, radar imagery indicated that the precipitation had weakened over Kansas City. At the same time, convergence weakened as the surface low weakened (Hales 1978).



The flooding that resulted from the 12 September storms was intensified as a result of the high moisture content and instability of the atmosphere. The existence of a short-wave trough also intensified the storm's flooding ability through the strengthening of upper level divergence. The low-level outflow boundary also contributed to the flood capacity of the storms. Finally, the existence of the nearly stationary frontal boundary, and the development of storms along the outflow boundaries of previous storms, greatly increased the chance that flooding would occur (Hales 1978).

## **2.6 Minnesota**

On 23-24 July 1987, Minneapolis-St. Paul experienced tornadoes, hail, and flash flooding. The Minneapolis-St. Paul International Airport received 254 mm (10 in) of precipitation in just twenty-four hours. This heavy rainfall, combined with intense rainfall from a few days earlier, led to the devastating flash floods in the area (Schwartz *et al.* 1990).

At the synoptic scale, a subtropical high pressure dominated the eastern two-thirds of the United States, while a strong low pressure existed over the Hudson Bay area. Several short waves moved through the northern United States, a normal event associated with evening summertime mesoscale convective systems (MCS). Two short waves were found: one existed over eastern Minnesota and western Wisconsin, associated with a mesoscale convective system; the other existed over Montana, recognized by the shifting of wind and vorticity maximum. Upward forcing dominated much of the troposphere over central Minnesota (Schwartz *et al.* 1990).

A large supply of moisture was available from the Gulf of Mexico, with precipitable water values over 150% of normal. The dewpoint temperatures were greater than 60 °F ahead of a southwest to northeast oriented cold front. By 12Z, the dewpoint temperatures at the surface had reached the 70's °F (Schwartz *et al.* 1990).

The mesoscale convective system that moved over northern Minnesota on the 22<sup>nd</sup>, created two outflow boundaries that contributed to the flash flooding, which occurred on the 23<sup>rd</sup>. The northern boundary developed into a squall line and mesohigh pressure system. The changes in wind and pressure marked the movement of this boundary. High theta-e values migrated toward the northeast as the mesohigh moved to the southeast (Schwartz *et al.* 1990).

The 00Z 24 July interpolated sounding for Minneapolis indicated the existence of very unstable air. The lifted index was calculated at -8 and the CAPE was calculated at 3,192 J/kg. Based on the CAPE value, it was determined that the system had the ability to create updrafts of more than 70 m/s. Convective storms developed parallel to the outflow boundary, which allowed for training of the storms and increased the chance for flash floods. Furthermore, the area along the outflow boundary west of Minneapolis-St Paul, where the highest CAPE occurred, would experience the most intense and rapid cell development (Schwartz *et al.* 1990).

At 0032Z 24 July, a strong line of thunderstorms stretched across northwestern Minneapolis and stratiform precipitation existed in the area of a mesohigh. Shortly after 0032Z, a tornado was reported approximately 30 miles to the north-northwest of the Minneapolis radar. By 0130Z, a mesolow developed at the convergence of the northern and southern outflow boundaries. The line of thunderstorms stretched over the radar site, while heavy rains fell along the boundary between newly developed cells and the stratiform rain, possibly due to a higher moisture content and weaker wind shear associated with the mesoscale convective system (Schwartz *et al.* 1990).

By 0310Z, a quasi-stationary train echo had developed, with the development of 12 new cells, from 0130Z to 0310Z. The MCS may have developed out of the westward expansion of the thunderstorms that were visible on satellite. Heavy rain continued to develop over Minneapolis as the cold front passed. The rain ended around 06Z (Schwartz *et al.* 1990).

Prior to the storm, on 22 July, weather statements had reported a good chance of severe weather *and* flash flooding over the Minneapolis area. However, later statements stressed only

the chance for severe weather, paying little attention to the chance for flash flooding. By 21Z on the 23<sup>rd</sup>, a tornado watch was issued, which further decreased the attention paid to the flash flood potential (Schwartz *et al.* 1990).

The number of other severe weather watches and warnings issued made forecasting flash floods for this event difficult. The forecasters had no time to properly analyze the data to determine the flash flood potential. To better prepare for similar events, forecasters need to already have a good idea of the weather patterns associated with local flash flooding (Schwartz *et al.* 1990).

### III. FLASH FLOOD CLIMATOLOGY AND FORECASTING

There are many complications associated with flash flood forecasting. Sometimes the conditions conducive to heavy precipitation do not exist at the time of the forecast (Doswell *et al.* 1996). Sometimes flash flooding coincides with other severe weather phenomena, such as tornadoes, which generally overshadow the threat of impending flash floods (Rogash & Smith 2000). Several methods have been developed to reduce the difficulties associated with flash flood forecasting. Various climatologies, both countrywide and regional, have been created to aid in the recognition of flash flood potential, while some places in the West that lack data have adopted alternative methods for identifying flash flood potential, such as lightning detection.

#### **3.1 Flash Flood Climatology**

Flash flood reports reflect the non-meteorological and meteorological factors that contribute to flash flooding. Non-meteorological factors include topography, vegetation, and the time of year the flash flood occurred. Meteorological factors include frontal movement and the location of the 500-mb ridges and troughs (Gaffin & Hotz 2000). From this data, Doswell *et al.* (1996) developed a climatology of storm types common to flash floods, whereas, Konrad (1997) identified common patterns associated with heavy rain events.

The majority of convective systems are multicellular. Multicell convection occurs when several cells develop relatively close together. A supercell, typically multicellular, usually possesses strong updrafts and high low-level moisture, both of which suggest heavy precipitation. However, supercells also tend to exhibit signs detrimental to the development of heavy precipitation. The main deterrent to heavy precipitation is the characteristic dry airmass in the lower midtroposphere, which suggests the existence of evaporation. Furthermore, high wind

speeds aloft suggest rapid system movement, which is not conducive to flash flooding (Doswell *et al.* 1996).

Mesoscale convective systems (MCSs) produce the majority of flash floods (Doswell *et al.* 1996), as well as much of the large amount of warm weather precipitation and severe weather frequently experienced throughout most of the United States (Stensrud & Maddox 1988). The MCSs typically distribute their convection in a linear pattern recognizable on radar. Doswell *et al.* (1996) explain that due to this linear pattern, the threat of a flash flood increases with the passage of each convective cell. Furthermore, these MCSs often occur in groups that interact with each other, increasing the challenge of forecasting associated weather phenomena (Stensrud & Maddox 1988). The lower levels of the atmosphere provide evidence of the existence of MCSs, whereas satellite data can provide evidence in the upper troposphere of the existence of convective systems (Fritsch and Maddox 1981).

A squall line, according to The Facts on File Dictionary of Weather and Climate, is a “line or band of active thunderstorms that produce strong winds and rain” (Smith 2001, 192). Doswell *et al.* (1996) divide squall lines into two types: squall lines seen by radar and squall lines seen by satellite. Squall lines visible on radar develop when a low-level outflow of precipitation-cooled air creates an updraft. Other lifting processes assisting in the development of squall lines, such as mountains, help force the outflow boundaries creating updrafts. Occasionally, MCSs are visible as a squall line on satellite, but not always on radar. This particular type of MCS does not usually produce flash floods. A nonconvective precipitation system can, however, produce flash floods as a result of forced uplift, such as produced by mountains (Doswell *et al.* 1996).

Convective cells can interact with each other to cause heavy precipitation. Some thermodynamic parameters associated with this type of interaction consist of high values of precipitable water, instability, low-level moisture, and high positive values of equivalent potential temperature. Positive vorticity advection associated with a short-wave trough at 500-mb tends to

exist with heavy rainfall. Low-level convergence associated with a frontal boundary also leads to heavy rainfall (Konrad 1997).

Heavy rain events commonly occur in an 850-mb ridge. It has also been found that heavy rain events occur along with moisture ridges at multiple levels. Instability and the presence of a lifting mechanism at the theta-e ridge also produce heavy rainfall (Konrad 1997).

Konard examined heavy rain events in the Southeast including the location of heavy rainfall, relative to the Appalachians. Heavy precipitation was found to occur at different locations depending on the orientation of the wind. From this research, Konard (1997) grouped heavy rain events in the Southeast into five patterns based on their synoptic properties.

Pattern 1 is characterized by a boundary convergence zone approximately 400 km to the northwest of the heavy rainfall. Southwesterly winds bring in warm, moist air from the Gulf of Mexico and are strongest near the heavy rainfall. These events occur throughout the warm season. The western region receives 30% of these events, 31% of which occur along the warm air ridge (Konrad 1997).

Pattern 2 is also characterized by a boundary convergence zone to the northwest, but close to the heavy rainfall. These events commonly occur along a slow-moving cold front that possesses more than one low-pressure center and exhibits high values for the 700-mb mixing ratio, K index, and the 200-mb divergence. Southwesterly winds bring in warm, moist air from the Gulf of Mexico and heavy precipitation occurs in the warm sector of a strong frontal system. This type of event is also characterized by a train echo precipitation pattern and usually occurs when the surface temperatures are cool and the 500-mb theta-e values are low (Konrad 1997).

Pattern 3 is characterized by a boundary convergence zone approximately 100 km to the southeast of the heavy rainfall. Weak winds and frontal overrunning are characteristic of this type of event. The highest values for the 700-mb mixing ratio and the K-index are located near the heavy rainfall. The Piedmont region experiences 36% of this type of heavy rain event (Konrad 1997).

Pattern 4 is characterized by low-level convergence from the heavy rain zone to the southwestern border of the study area. The eastern and southern regions experience over 30% of this type of event. The western zone experiences only 17%, as a result of the rainshadow effect caused by the southeasterly winds. Furthermore, these events only tend to occur during the months from June through August (Konrad 1997).

Finally, pattern 5 is characterized by two regions of low-level convergence. Although this pattern tends to have no well-defined surface pattern, several events in the study experienced a weak front that was oriented northwest to southeast, typically in the frontolysis stage. Over 75% of these events occur between June and August (Konrad 1997).

Several other types of climatologies have been created to aid in flash flood potential recognition. Maddox *et al.* (1979) created a climatology of four event types associated with flash flooding in the United States from 1973 to 1977: Synoptic, frontal, mesohigh, and western. Maddox *et al.* (1980) also added a more extensive climatology of patterns for flash flooding in the West.

Synoptic events consist of a quasi-stationary surface front and a major 500-mb trough generally occurring during the spring and fall (Maddox *et al.* 1979). In the southern Appalachians, precipitation occurs ahead of a southwest to northeast oriented front. This event type also experiences southerly winds at the surface and southwest winds above 850-mb (Gaffin & Hotz 2000).

Frontal events are associated with heavy rains caused by warm, unstable air flowing over a west-to-east oriented quasi-stationary frontal boundary (Gaffin & Hotz 2000; Maddox *et al.* 1979). These events tend to occur during the night, and the majority of them have an associated meso- $\alpha$  scale short-wave trough (Maddox *et al.* 1979). However, the large-scale patterns are weak and the flash flooding occurs near the 500mb ridge. This event type also experiences southerly winds at the surface and southwest winds above 850-mb. Furthermore, precipitation occurs behind the surface front (Gaffin & Hotz 2000).

The most common type of flash flood event is a mesohigh, which results from prior convective activity. This event type is associated with heavy precipitation near the 500-mb large-scale ridge (Maddox *et al.* 1979), while westerly winds are common from the surface to 500-mb. The mesohigh events are also characterized by a quasi-stationary front, wherein flash flooding occurs due to cold air outflow boundaries, resulting from previous storms. These flash floods occur to the southwest of the center of a meso-high (Gaffin & Hotz 2000).

Western events are generally associated with “weak large-scale patterns without well-defined surface patterns” (Maddox *et al.* 1979, 122). These events occur during the afternoon and early evening because of a very moist troposphere and an unstable atmosphere. These flash flood events are placed into one of four categories, based on each event’s 500-mb patterns (Maddox *et al.* 1980).

The Type I category is associated with a weak short-wave trough on the western side of a synoptic ridge. These events often develop along a frontal boundary beneath the synoptic ridge, and are associated with high surface dewpoints and very moist conditions as high up as 300-mb. The precipitable water value is 23.9 mm, a value 184% of the average. The K-index has an average value of 40, while the Lifted index has a value of -4, both of which indicate instability. Type I events typically occur from June through August over much of the West (Maddox *et al.* 1980).

Type II events are associated with a short-wave trough on the eastern side of a synoptic ridge. The surface temperatures are warm and the atmosphere is moist through the 300-mb level. On average, the precipitable water value is 25.4 mm, 145% of the normal, and the K-index and the Lifted index are 39 and -5, respectively. These Type II events usually occur from July through September over the southeastern portion of the West (Maddox *et al.* 1980).

The Type III category is characterized by strong synoptic patterns, which affect large areas and develop as a result of orographic lifting. Satellite data is necessary to provide a visual account of the storm’s development, since it moves in from the Pacific Ocean. The precipitable



water is 18.3 mm, 159% of the average. The K-index and the Lifted index are 27 and +1 respectively. The Type III events affect the Pacific coast from December through March (Maddox *et al.* 1980).

Type IV events are characterized by a synoptic zonal pattern. These events tend to coincide with a weak mid-level short-wave trough, and often accompany the southwest monsoon. The Type IV events occur during July over New Mexico and Arizona (Maddox *et al.* 1980).

### **3.2 Flash Flood Forecasting**

Forecasters do not always recognize the signs indicating the potential for flash flooding, due to the fact that rain is very common and usually harmless (Doswell *et al.* 1996). Furthermore, the difficulties associated with flash flood forecasting are increased when other challenging factors are present, such as accompanying severe weather (i.e. Tornadoes) and the affects of complex terrain.

#### *a) Tornadoes and Flash Floods*

Severe weather phenomena like tornadoes can cause meteorologists to overlook the signs of impending flash floods (Rogash & Smith 2000). Even though flash floods and tornadoes cause most of the fatalities and much of the property damage associated with thunderstorms (Rogash and Racy 2002), flash flooding remains the leading fatal convective storm event annually (Doswell *et al.* 1996). In fact, from 1955 to 1995, the National Climatic Data Center (NCDC) recorded an annual average of 73 fatalities associated with tornadoes, compared to 136 fatalities associated with flood events (Rogash and Racy 2002).

One such storm occurred in early June 1980, when tornadoes and flash floods accompanied severe thunderstorms as they traveled through Nebraska, Illinois, and Indiana (Maddox 1981). Maddox (1981) reported that 127-178 mm (5-7 in) of rain fell over Nebraska

and Missouri causing flash flooding. The surface pattern of this system resembled the pattern associated with “‘mini-outbreaks’ of tornadic storms” (Maddox 1981, 1583).

Maddox and Doswell (1982) analyzed three storm events, which demonstrated that the upper and low-level jet characteristics are not always clear. The forecasts for each event expected intense thunderstorms to occur in regions of neutral or positive vorticity advection at 500-mb. Although the characteristics of these storms were not typical for convective severe weather, they were common for heavy rain and flash flooding, which all of the storm events experienced (Maddox & Doswell 1982).

Although vorticity advection is not always dependable, low-level thermal advection can help identify areas of low-level convection. Warm advection is common at the 700-mb level for nocturnal thunderstorms. Tornadic activity and flash flooding also commonly occur in areas associated with warm advection. Maddox and Doswell (1982) suggested that when the 850-mb warmest temperature axis lies west of the 850-mb moisture axis, severe storms are most likely to occur. During the 3 April 1974 tornado outbreak, severe weather occurred in an area of low-level warm advection and significant upward motion. The following events developed in areas of strong warm advection and conditional instability (Maddox & Doswell 1982).

On 11-12 April 1980, tornadoes, hail, and heavy rains occurred over northeastern Texas, southern Arkansas, and parts of Mississippi. At 12Z on the 11<sup>th</sup>, there was a westerly flow at the 200-mb level with an area of wind speeds between 70 and 85 knots. Thunderstorm activity occurred in the area where the upper and lower level jets intersected. Strong vertical wind shear dominated this area. However, a clear jet streak was absent (Maddox & Doswell 1982).

By 00Z on the 12<sup>th</sup>, convection was present on both the east and west sides of the lower-level jet. Weakening of the upper-level flow and an increase in diffluence had occurred. The vertical wind shear had also weakened in the area of thunderstorm activity. The Limited-area Fine-mesh Model (LFM) had forecasted the existence of intense short-wave troughs and vorticity

centers northwest of the area in which thunderstorms were actually occurring (Maddox & Doswell 1982).

By 12Z on the 12<sup>th</sup>, the upper-level jet streak had changed dramatically. The maximum wind speeds exceeded 130 knots, while the lower-level jet had weakened and split. Thunderstorms continued to the west of the lower-level jet over the Mississippi Valley, while new thunderstorms developed over southern Texas. In the area of thunderstorm activity, the upper-tropospheric winds weakened while strong diffluence continued (Maddox & Doswell 1982).

On 2-3 July 1980, severe thunderstorms, tornadoes, hail, and high winds swept through central Missouri and Kentucky. Indiana, Kentucky, Illinois, and Missouri experienced flash flooding associated with heavy rain. At 12Z on the 2<sup>nd</sup>, strong convective storms existed beneath the 200-mb ridge, near the lower-level jet and approximately 1,000 km south of the upper-level jet streak. Relatively weak vertical wind shear dominated this area (Maddox & Doswell 1982).

While the main storms moved across Missouri and Illinois, new storms developed over Kentucky and Indiana, and by 23Z, these two systems had merged. By 00Z on the 3<sup>rd</sup>, the storms remained under the upper-level ridge where the vertical shear remained weak. The storms that developed over the Ohio Valley were developing in a region considered unfavorable for storm development. The LFM had forecast that the storms would develop in an area where the vorticity advection was neutral or positive. At 12Z, the storms remained in the upper-level ridge, which continued to exhibit weak vertical wind shear (Maddox & Doswell 1982).

#### *b) Forecasting Flash Floods in Complex Terrain*

Flash floods tend to occur in mountainous terrain due to the “prevalent short response times of watersheds” (Warner *et al.* 2000, 797). Orographic lifting also tends to ignite heavy precipitation, resulting in flash flooding (Warner *et al.* 2000; Maddox *et al.* 1978). As a result, forecasters encounter difficulties when forecasting for complex terrain (Maddox *et al.* 1995).

Changes in landuse/landcover affect the ability of the soil to absorb water, and can exacerbate flash flooding. For example, in May 1996, a wildfire burned 50 km<sup>2</sup> within the Buffalo and Spring Creek watersheds in Colorado. Two months later, on the 12<sup>th</sup> of July, a weak front moved in from the northeast causing the winds to blow upslope. To the west of the front, dewpoint temperatures were between 55°F and 65°F. However, the only convection that existed near Colorado during that morning was in Oklahoma, moving southward. Convection over the mountains of Colorado did not occur until around noon. During the storm that ensued, more than 80 mm (3.2 in) of precipitation fell over the charred area within a single hour, flooding the watersheds of Buffalo Creek and Spring Creek, where the forest had been destroyed (Chen *et al.* 2001).

Several mountain ranges affect Arizona's climate, and complicate forecasting. Daily thunderstorms commonly develop during the wet monsoon season, and the changes in elevation from southwest to northeast enable the wet monsoon storms to create heavy precipitation and flooding (Maddox *et al.* 1995). Weather radar, a useful tool for flash flood prediction, enables the estimation of precipitation, which provides a means to reduce the risk of damage and injury related to flash floods. Unfortunately, radar blockage caused by the mountainous terrain can lead to inaccurate precipitation estimates (Vivekandan *et al.* 1999). To complicate matters, the radar coverage in Arizona is incomplete, further inhibiting the forecast of flash floods (Holle & Bennett 1997).

To compensate for the lack of radar data, the frequency of cloud-to-ground lightning strikes has been implemented by the National Weather Service (NWS) in the West for more than a decade. A relationship between lightning strike frequency and flash flood occurrence exists for flash flood events, such as the Big Thompson Storm in 1976. A flash flood that occurred in Ohio in June 1990 also illustrated that lightning strike frequency can provide valuable information as to the location and movement of flash flood producing systems. For instance, on 7 July 1990, an Arizona flash flood accompanied cloud-to-ground lightning strikes. Lightning flashes were

recorded, beginning approximately three hours before flash flooding was reported. At the time the flash flood occurred, reports of the majority of the lightning flashes ceased (Holle & Bennett 1997).

In fact, this phenomenon occurred with the majority of the flash flood events examined by Holle and Bennett. Lightning flashes generally occurred within ten hours of the flash flood event. However, few to no lightning flashes were recorded after the flash flood occurred. Only one of these flash flood events saw a high lightning flash frequency occur more than eight hours before the flash flood event, and only one saw lightning flashes over an hour after the flash flooding occurred (Holle & Bennett 1997).

Examination of the relationship between lightning flashes and flash flooding found that the lightning flash frequency provided no indication of when a flash flood would occur or of the strength of the flash flood. However, lightning flashes did occur more frequently on days that experienced a flash flood than days when no flash flood occurred. Furthermore, days that did not experience flash flooding were most likely to experience lightning flashes at low frequencies (Holle & Bennett 1997).

Two flash flood events were examined to determine the relationship between lightning flashes and streamflow: 24 July and 19-20 July 1990. On the 24<sup>th</sup>, 36 mm of rain fell over the upper tributaries of the watershed northeast of Phoenix. The area east of Tucson received rain gauge totals of 16-89 mm. Between 0400 and 1100 MST, 553 lightning strikes were recorded during the flash flood. Comparison of the relationship between lightning flashes and streamflow show that streamflows were at about zero until the lightning flash frequency began to decrease. Between 0600 and 0900 MST, the development of a large mesoscale convective system occurred (Holle & Bennett 1997).

The flash flood event of 19-20 July experienced two separate periods of lightning and rain. The first period was more active than the second. Rainfall amounts were recorded at an average of 28 mm northeast of Tucson. Nine hundred and ninety-three lightning flashes were

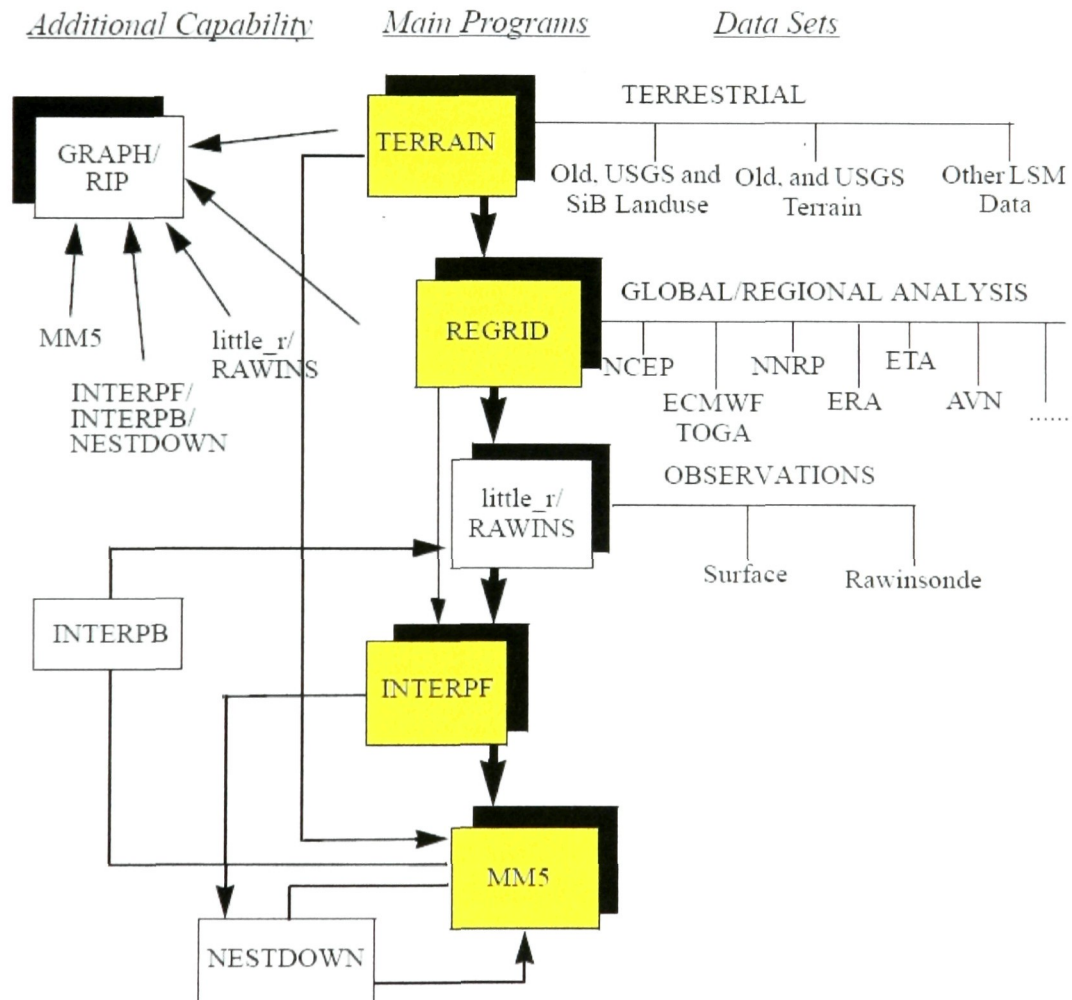
recorded over the eastern region of the study area. Rainfall amounts in the area were recorded between 18 and 39 mm, and the majority of the recorded precipitation occurred during periods of more frequent lightning flashes. In fact, the streamflows also increased during the lightning periods, but decreased to zero *between* the lightning periods (Holle & Bennett 1997).

Since higher frequencies of lightning flashes are associated with flash flood occurrences, an understanding of the relationship of lightning flashes and Eastern Kentucky flash floods could provide another tool with which forecasters could predict the most probable locations for flash flood occurrence. As seen in the Arizona cases (July 1990), lightning flashes occurred in the same general location as the flash floods.

Obviously, many variables affect the accuracy of flash flood forecasting. Forecast models, such as MRF, NGM, and ETA predict the movement and development of weather phenomena. These models can prove useful by providing data for areas that lack data, improving the accuracy of forecasts.

#### IV. SIMULATING STORM EVENTS

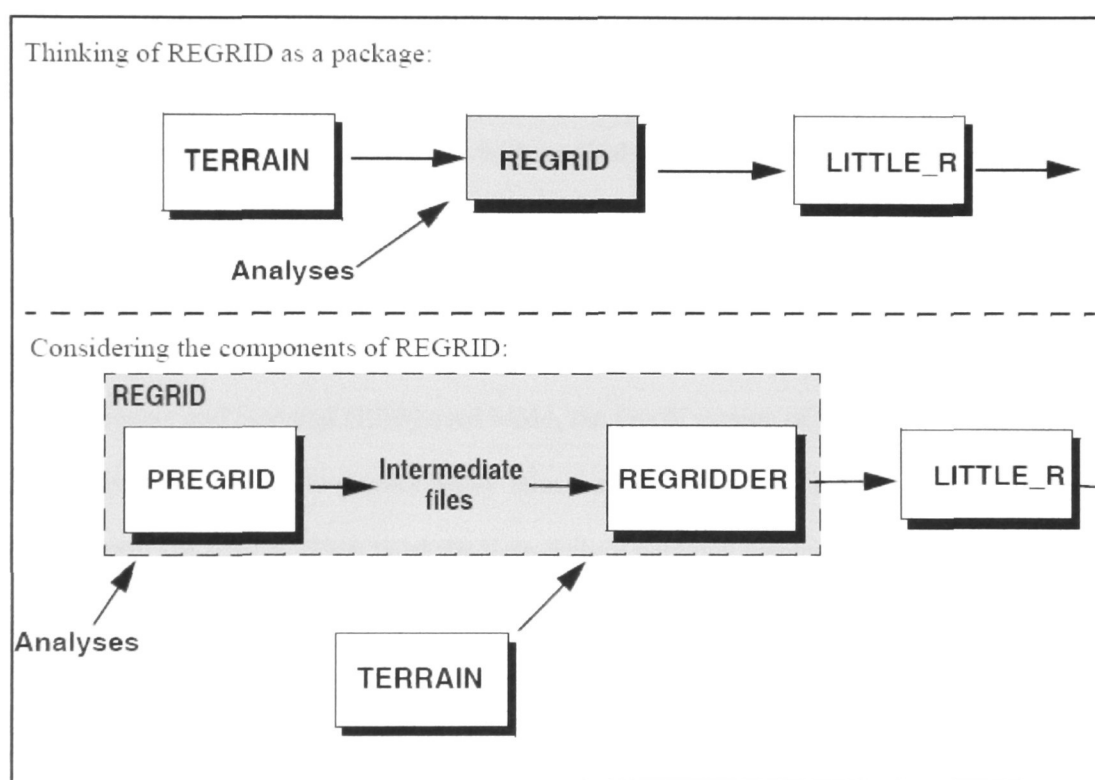
Numerical models simulate the lifecycles of weather systems, which help forecasters make more accurate predictions. One such model, the PSU-NCAR Mesoscale Model, MM5, is the fifth generation of a numerical weather prediction model originally used at Pennsylvania State University during the early 1970s. This mesoscale model consists of a number of programs, including Terrain, Regrid, Interpf, and MM5 (Figure 4.1; Dudhia et al. 2005).



**Figure 4.1: Model of the MM5 System (Dudhia et al. 2005)**

The first program in the MM5 sequence, Terrain, identifies geographic details. Terrain essentially converts the surface data, such as the latitude-longitude coordinates of the model area and the characteristics of the land, into the format used by MM5 (Dudhia et al. 2005).

Regrid consists of two programs: Pregrid and Regridder (Figure 4.2). Pregrid acquires meteorological analyses and forecasts from a file. Regridder utilizes the data from Pregrid as well as the terrain, map, and landuse data from Terrain. These data are then transformed into the MM5 grid (Dudhia et al. 2005).



**Figure 4.2: Model of Regrid (Dudhia et al. 2005)**

Regrid then sends the data to Interpf. Here, the pressure levels are converted into the sigma coordinates used by MM5. Interpf then uses the data to create the initial conditions for the model. Based on the data obtained from Interpf and the parameters chosen in the MM5 program, MM5 makes a prediction (Dudhia et al. 2005).



One of the parameters, the cumulus parameterization scheme, identifies the processes MM5 will use in the development of clouds. There are several cumulus parameterization schemes. Some are designed to be most effective with a grid size greater than 30 km, such as Anthes-Kuo, Arakwa-Schubert, and Betts-Miller. Others work best with a grid size less than 30 km, such as Grell (10-30 km) and Kain-Fritsch (20-30 km). The option, None, works best under 5-10 km (Dudhia et al. 2005).

MM5 consists of four other major physics schemes: Microphysics (Explicit Moisture), Planetary Boundary Layer, Radiative, and Surface. The Microphysics Scheme deals with precipitation processes, such as how ice and water behave around the freezing line. The Planetary Boundary Layer Scheme deals with vertical mixing processes in the boundary layer. The Radiative Scheme determines the interaction of short-waves and long-waves with the surface and atmospheric moisture, whereas the Surface Scheme distinguishes between land and water surfaces (Dudhia 2005).

Spencer and Stensrud (1998) used MM4, the fourth version of the model, to examine the effectiveness of the model to predict heavy rains, capable of producing flash floods, over areas with relatively flat terrain. They used the Kain-Fritsch cumulus parameterization scheme for the inner domain, and the Anthes-Kuo cumulus parameterization scheme for the outer domain. Without any modifications, the model indicated less precipitation than observed in both amount and areal extent. However, the area indicating the highest precipitation corresponded relatively well to that observed (Spencer & Stensrud 1998).

To improve the simulation's accuracy, three alterations were implemented. Since the model failed to simulate enough precipitation, an adjustment was made to maximize the precipitation efficiency. This adjustment increased the area receiving precipitation and the amount simulated. In fact, some of the storms analyzed actually simulated more precipitation than observed. Another modification eliminated downdrafts below the cloud base. This modification greatly increased the area and amount of precipitation, as well as the period of

precipitation, all of which were greater than observed. The final modification created a delayed downdraft, resulting in a smaller area and amount of precipitation simulated, matching more closely the observed data. This study indicated that simulating heavy rain in relatively flat terrain is sensitive to downdrafts (Spencer & Stensrud 1998).

Another simulation used MM5 to analyze the Super Outbreak of 3-4 April 1974, during which one hundred and forty-eight tornadoes were reported in thirteen states. A previous study had indicated the existence of a cold front aloft that might have helped spark the intense storms during this event. Furthermore, the detached nature of the thunderstorm activity from the frontal boundary also supported the existence of a cold front aloft (Locatelli *et al.* 2002).

Locatelli *et al.* (2002) applied MM5 to simulate the atmospheric conditions associated with the Super Outbreak to determine how the cold front aloft may have affected the production of severe weather. To accomplish this goal, the Kain-Fritsch cumulus parameterization scheme was implemented for two domains: a 54-km resolution outer domain and an 18-km resolution nested domain. The model represented the main atmospheric conditions well, although the model analyzed the major cold front as an occluded front, and failed to represent the northern most line of storms. Nevertheless, the model accurately placed the precipitation zones with respect to the fronts. The existence of a cold front aloft was confirmed, accompanying the central line of storms, which produced the majority of the severe weather (Locatelli *et al.* 2002).

Clearly, MM5 is a useful tool for storm prediction, even if it does not always accurately predict the specifics associated with the storm event. Nevertheless, MM5 does help to identify areas most likely to experience severe weather, which, combined with knowledge of the threat area, can greatly assist in producing accurate forecasts. Furthermore, studying past storms with MM5 can help improve the accuracy of the model, as well as the information interpreted from it.

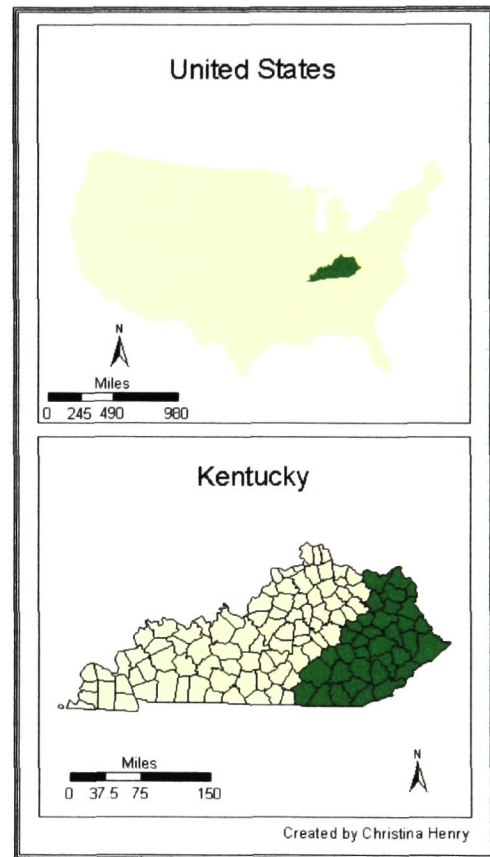
## V. STUDY AREA, DATA, AND METHODOLOGY

### 5.1 Study Area

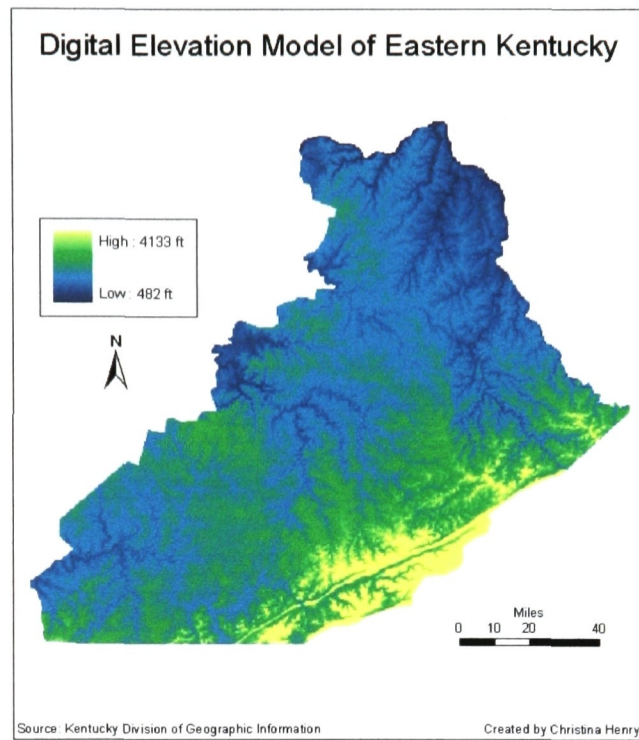
Kentucky lies in the eastern United States, bordered by the Ohio River to the north and the Appalachians to the east (Figure 5.1). The elevation of eastern Kentucky slopes downhill from the southeast to the northwest (Figure 5.2). Black Mountain, located in Harlan County, represents the highest point in southeastern Kentucky, at 4,145 ft; (O'Dell 2006).

Except for the Appalachians to the east, the lack of major physical boundaries allows for an influx of cold air from the north and warm moist air from the southwest. Consequently, Kentucky receives the majority of its moisture from the Gulf of Mexico (Ulack 1998).

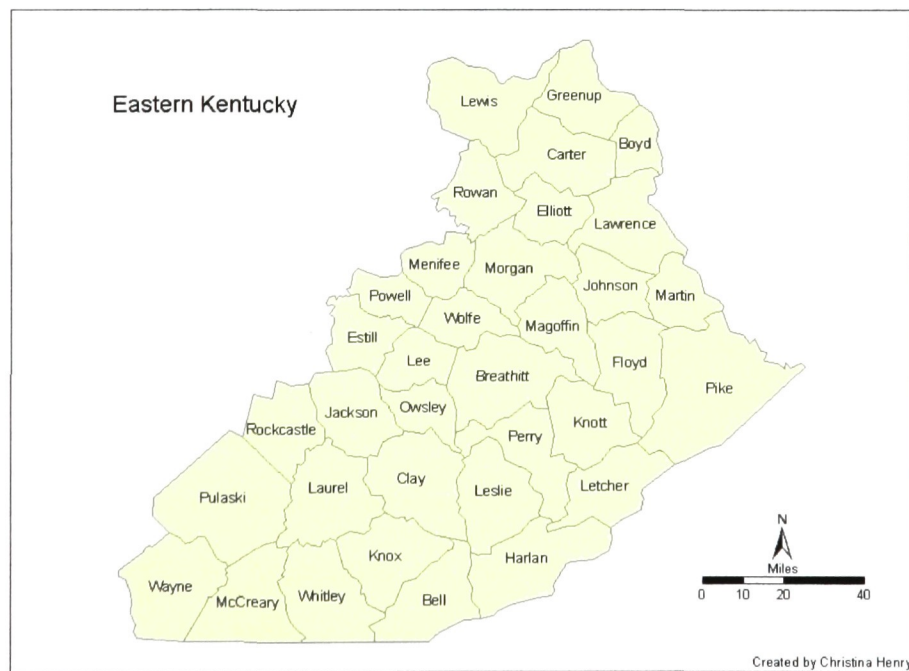
Kentucky has a humid subtropical climate, with average temperatures of 53 °F in the northeast to 59 °F in the southwest. The annual precipitation ranges from 1,066.8 mm (42 in) in the north to 1,320.8 mm (52 in) in the south (Foster 2006). The state is divided into four climate divisions: Western, Central, Bluegrass, and Eastern. This research focuses only on the Eastern Climate Division, hereafter referred to as Eastern Kentucky (Figure 5.3). Of these thirty-five counties, the NWS offices in Charleston West Virginia and



**Figure 5.1: Eastern Kentucky is located in the foothills of the Appalachians.**

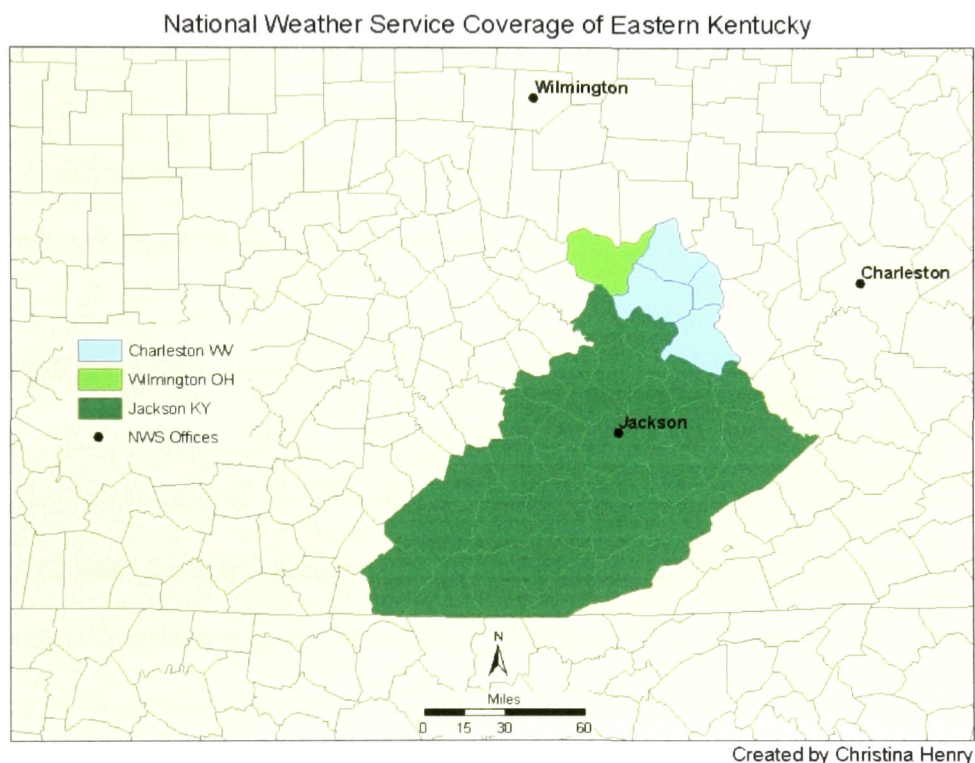


**Figure 5.2: Digital Elevation Model of Eastern Kentucky**



**Figure 5.3: Eastern Kentucky is defined by the boundaries of Kentucky's Eastern Climate Division**

Cincinnati Ohio cover the five northern most counties. The NWS in Jackson, Kentucky covers the remainder of this region (Figure 5.4). This office actually began operations in 1981 as a result of a particularly devastating flash flood event that occurred 4-6 April 1977, when 152.4-228.6 mm (6-9 in) of precipitation resulted in six deaths (NWS Jackson 2005; NOAA 1977).



**Figure 5.4: County Warning Areas for Eastern Kentucky**

## 5.2 DATA

The remainder of this research focuses on the climatology associated with flash flooding in Eastern Kentucky from 1993-2002 and examines the atmospheric conditions associated with the flash flood event of 3-4 August 2001. A local climatology was created from flash flood reports obtained from the National Climatic Data Center's (NCDC) online Storm Reports Database, and precipitation data was obtained from the Midwestern Regional Climate Center's (MRCC) online database, MICIS.

Surface and upper air data (including sounding data for Nashville TN and Wilmington OH) were obtained from Plymouth State College (PSC). The upper air data were available for every 12 hours, 00Z and 12Z, while the surface data were available every hour. Radar data for the entire Mid-Atlantic region were also obtained from PSC. NEXRAD level II and III radar data were obtained from NCDC and were available for every 5-6 minutes. Forecast discussions and warning reports were obtained from the National Weather Service (NWS) office in Jackson, Kentucky. Satellite and surface analysis charts were obtained from UNISYS.

Unfortunately, local data for Eastern Kentucky is limited. Therefore, MM5 simulations were conducted to provide data for the local scale.

### **5.3 METHODOLOGY**

#### *a) Creating a Climatology*

The first step in understanding the nature of a flash flood event is to examine the local climatology. Therefore, precipitation and flash flood climatologies were constructed for Eastern Kentucky. Unfortunately, this task proved difficult due to the incomplete data available.

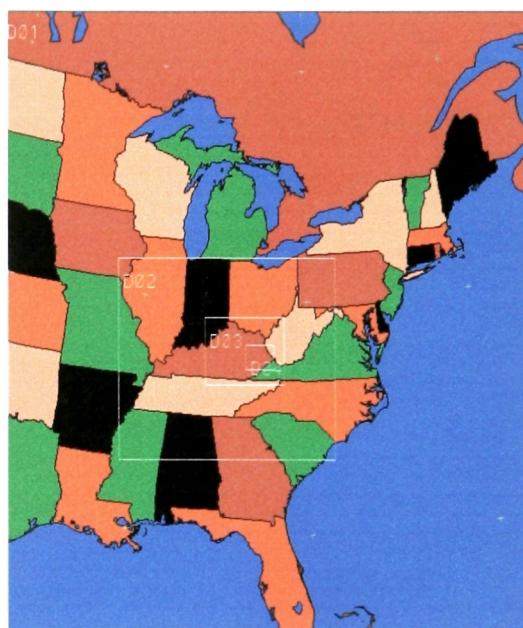
To create a precipitation climatology by county and season, the number of days receiving 25.4 mm (1 in) of rain or more were counted for each station in Eastern Kentucky, that had records, from 1993-2002. Some counties had no representation while others had multiple stations. In order to minimize the inaccuracies caused by the disproportionate number of stations, the data was normalized by dividing the average precipitation days by the number of stations in the each county. The data were then recorded and mapped in Arcmap. A flash flood climatology was then created by counting the number of flash floods reported in each county, by season, from 1993-2002. This data was also mapped in Arcmap.



*b) Simulating the Storm*

Surface and radar data are the only observed meso-scale data available for eastern Kentucky. However, forecasting flash floods requires knowledge of the mesoscale upper air conditions as well. Therefore, MM5 simulations provided high spatial-temporal data, for both surface and upper air conditions, and helped clarify the meso-scale setting for the current flash flood event.

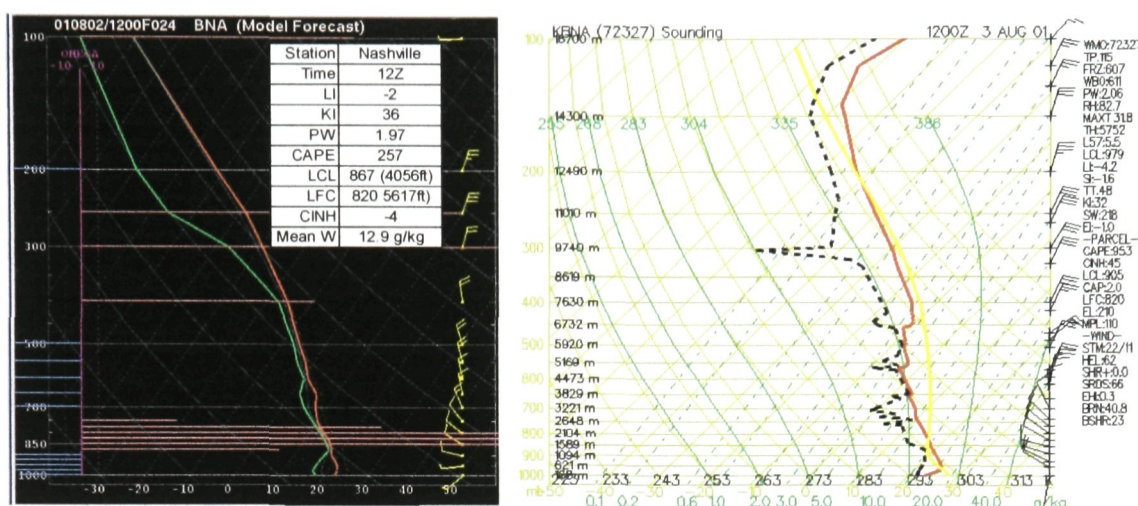
Four simulations were conducted using four domains each (Figure 5.5). The simulations were initialized using NCEP-NCAR data and ran from 12Z August 2<sup>nd</sup> through 06Z August 5<sup>th</sup>, utilizing thirty-eight sigma levels. These simulations used five main physics parameters, including the MRF Planetary Boundary Layer Scheme and the Cloud Radiation Scheme. The Simple Ice Explicit Moisture indicated ice melt immediately below the freezing line. The Surface Scheme consisted of five soil layers, which allowed for variation in ground temperature (Dudhia et al. 2005). Only the



**Figure 5.5: MM5 Domains – Domain 1 has a resolution of 27 km, Domain 2 has a resolution of 9 km, Domain 3 has a resolution of 3 km, and Domain 4 has a resolution of 1 km.**

Cumulus Parameterization Scheme was altered for each simulation.

The first simulation utilized the Grell Cumulus Parameterization Scheme for all four domains (Grell et al. 1994), whereas the second simulation used Kain-Fritsch for all four domains (Kain & Fritsch 1998). These two schemes were chosen because they work best at the smaller scales. The other two simulations continued to use Grell and Kain-Fritsch, respectively, in the outer two domains, but switched to the None Scheme for the inner two domains. The None Scheme was chosen because it works best at the smallest scales.



**Figure 5.6: The Modeled (Left) and Observed (Right) Sounding for Nashville TN at 0800 EDT (12Z) 3 August 2001**

Once the simulations were complete, a qualitative comparison was made between observed and modeled soundings for Nashville, Tennessee and Wilmington, Ohio. A visual comparison was also conducted for the simulated and the observed radar reflectivity, using both radar data from PSC and the level III reflectivity data for Jackson, Kentucky.

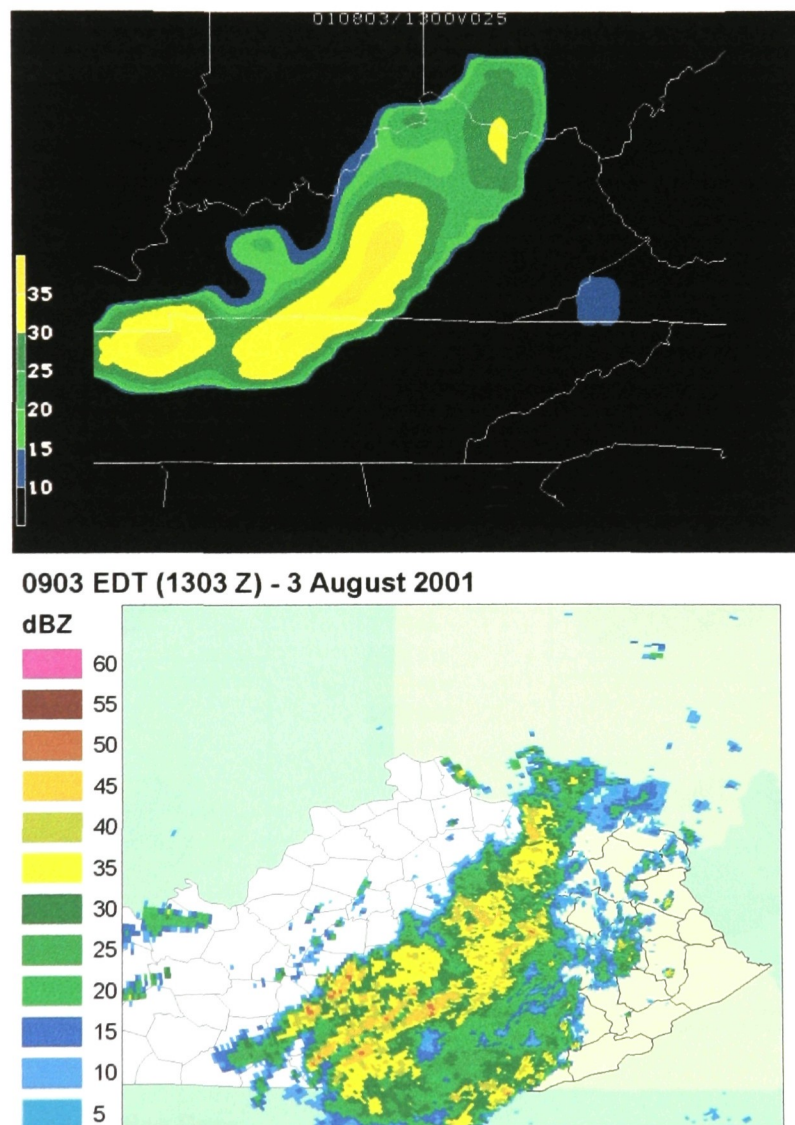
The Grell soundings matched the observed soundings relatively well. Both Wilmington and Nashville, observed and modeled, indicated a dry middle layer, marginal instability, and westerly winds. However, the observed soundings for both stations indicated larger directional wind shear than the modeled soundings.

The visual analysis of the observed and modeled radar reflectivity also indicated that the Grell simulation provided the most satisfactory results. The model reflectivity matched relatively well, although the model's intensities lagged the observed reflectivity by as much as two hours. However, the model reflectivity was less accurate with respect to the location of observed reflectivity; the model needed to be advanced by five to six hours to match the actual observed location. Furthermore, the observed reflectivity tended to indicate heavier precipitation than the model reflectivity until around 0800 EDT (12Z) 3 August. Afterwards, the model tended to indicate heavier precipitation than the observed. In addition, the model reflectivity was more



elongated than the observed reflectivity throughout the event. The other simulations provided smaller areas and smaller intensities of reflectivity, which were less accurate than the Grell simulation.

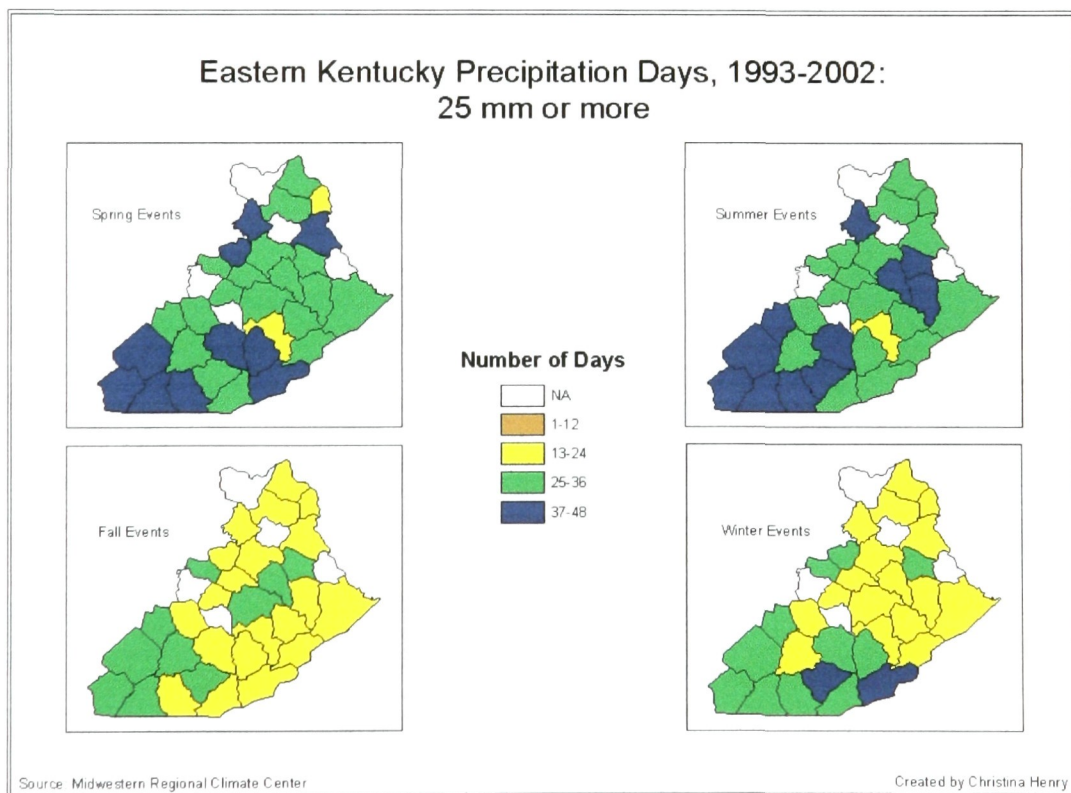
The simulation using the Grell Cumulus Parameterization Scheme for all four domains provided the most accurate results. Therefore, the upper-air data obtained from the Grell simulation was used to represent the atmospheric conditions in the local storm analysis (Chapter 7.2).



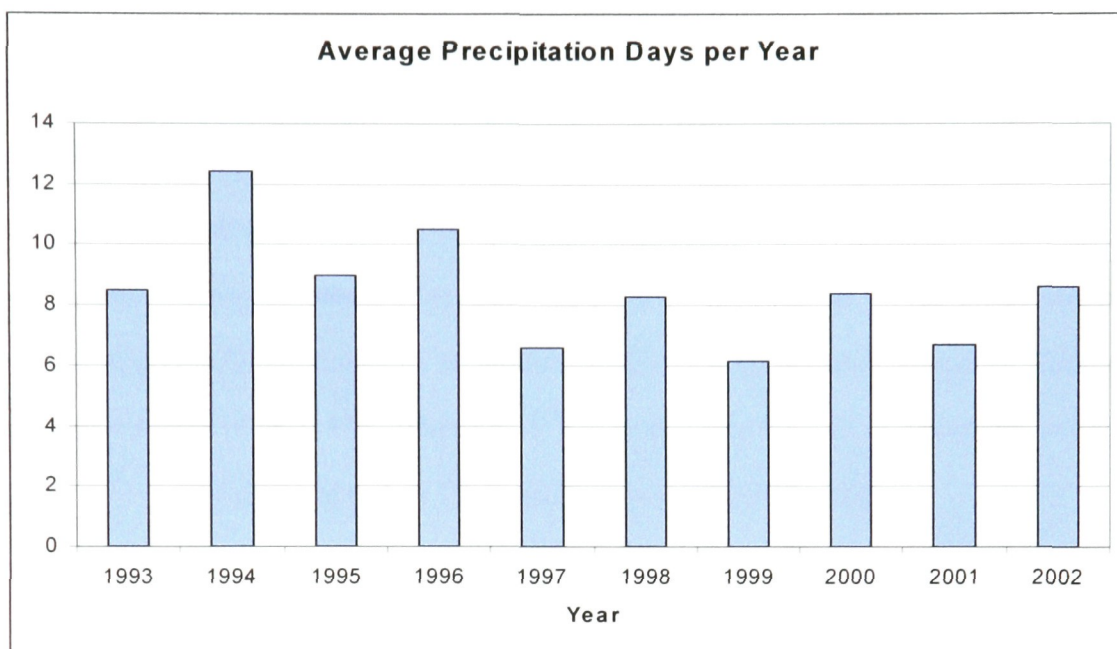
**Figure 5.7: The Modeled (top) and Observed (bottom) Radar Reflectivity for 0900 EDT (13Z) 3 August 2001.**

VI.  
EASTERN KENTUCKY PRECIPITATION AND FLASH FLOOD CLIMATOLOGY

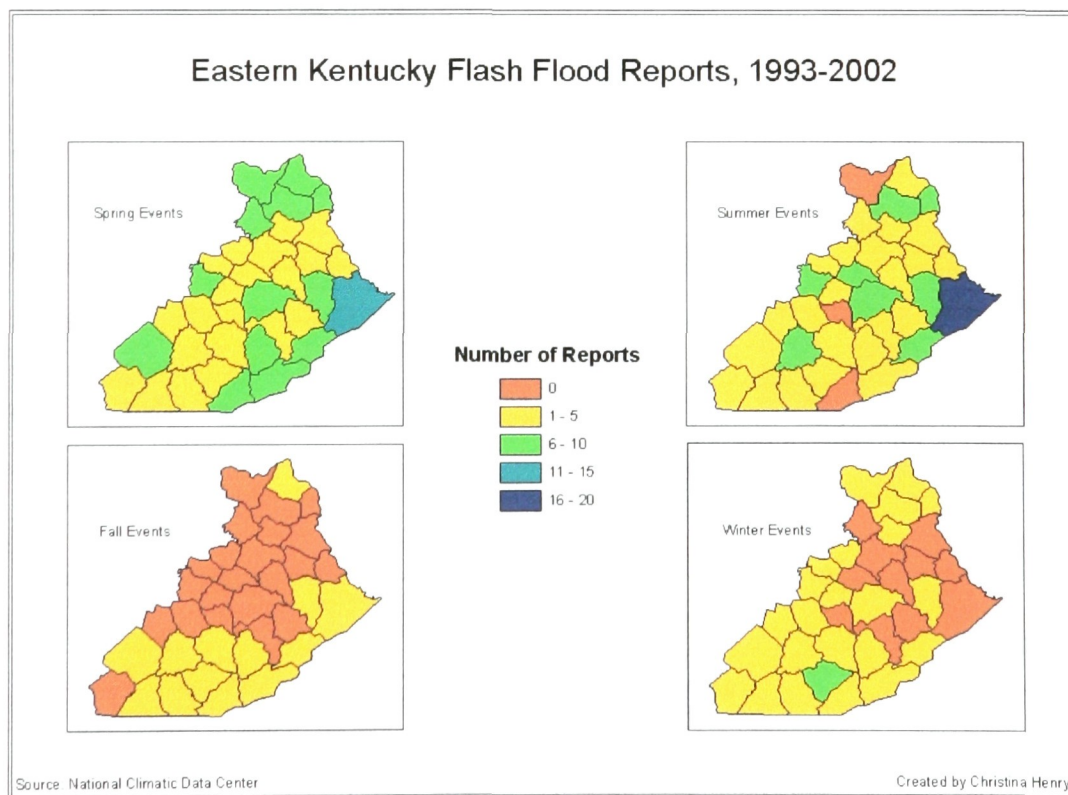
Based on 1895-2003 data, Eastern Kentucky receives an average of 1,192 mm (46.42 in) of precipitation per year. From 1993 to 2002, however, this area received a higher annual average of 1,238 mm (48.76 in), indicating wetter than normal conditions (MRCC 2004). During this decade, Eastern Kentucky experienced an average of 9 days per year in which at least 25.4 mm (1 in) of precipitation was recorded (hereafter referred to as precipitation days; Figure 6.1-6.2). On average, each precipitation day received 38 mm (1.5 in) of rain.



**Figure 6.1: Precipitation Days**



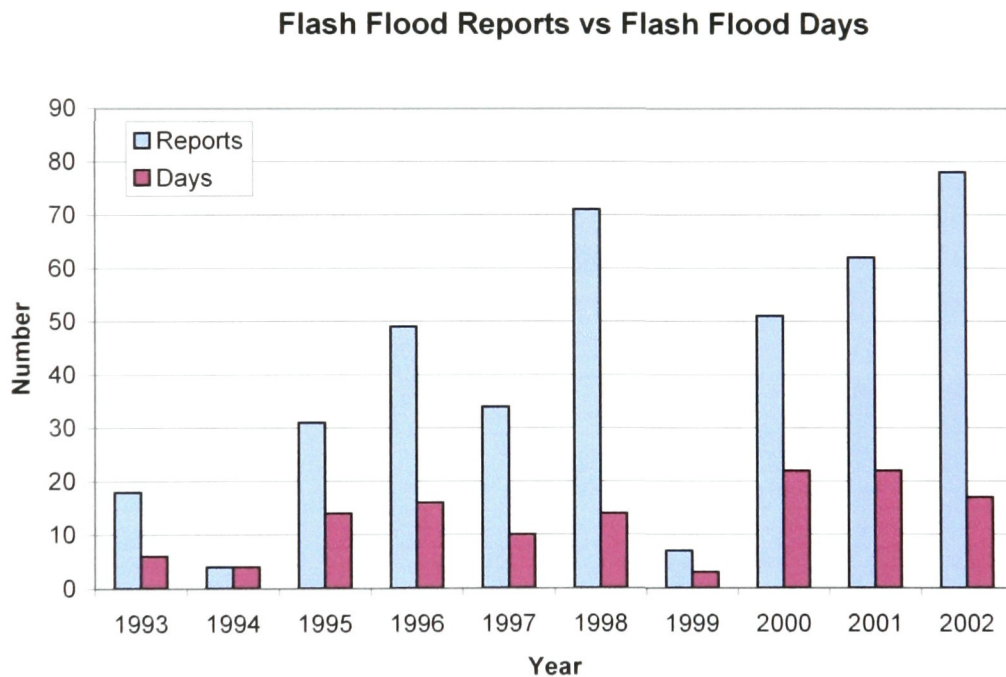
**Figure 6.2: Precipitation Days per year for Eastern Kentucky**



**Figure 6.3: Flash Flood Reports**

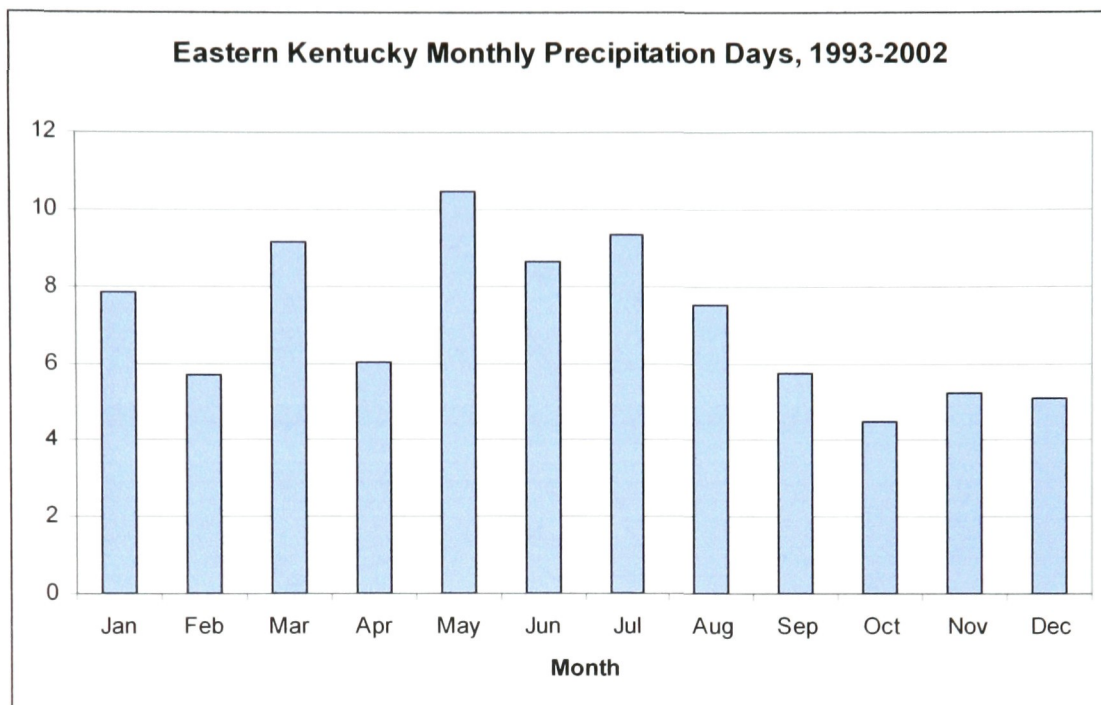
Between 1993 and 2002, 405 flash floods were reported in Eastern Kentucky (Figure 6.3), with an average of 41 flash floods per year. Annually, an average of 13 days experienced flash flooding, with approximately three flash floods per day (Figure 6.4). Of these 405 flash floods, 14 caused at least \$1 Million worth of property damage.

The majority of precipitation days and flash floods occurred during the spring and summer (Figure 6.5). Furthermore, these flash floods primarily occurred during the afternoon and evening hours of 15-23 EDT (Figure 6.6). Gaffin and Hotz (2000) found similar records for the southern Appalachians of eastern Tennessee and western North Carolina.

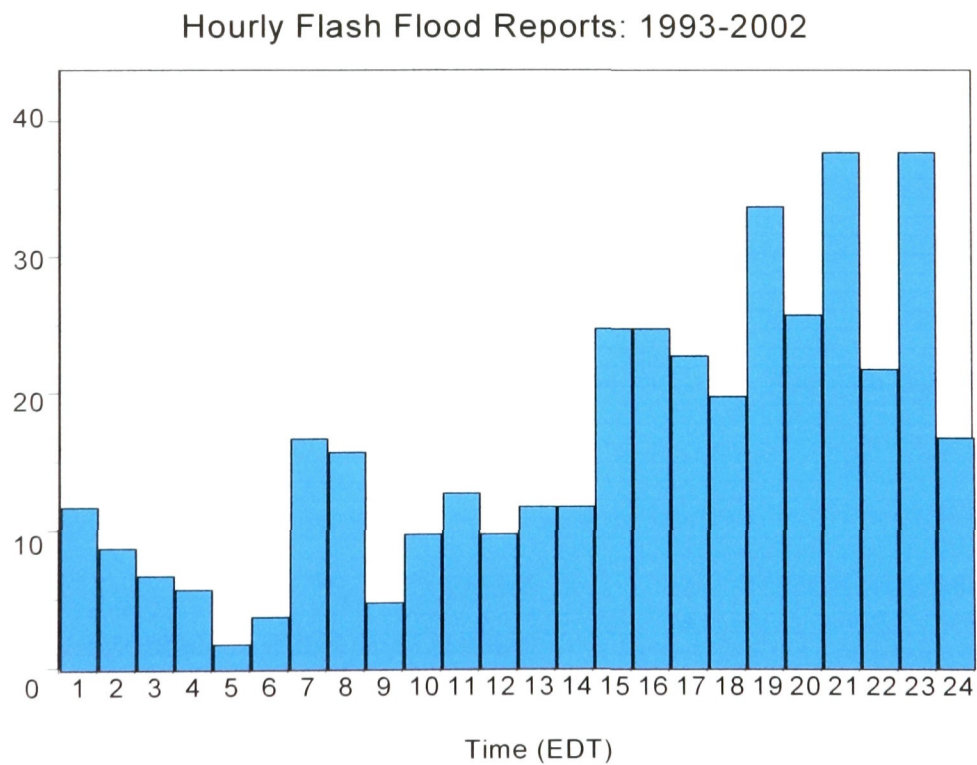


**Figure 6.4: Annual Flash Flood Reports vs. Annual Flash Flood Days**

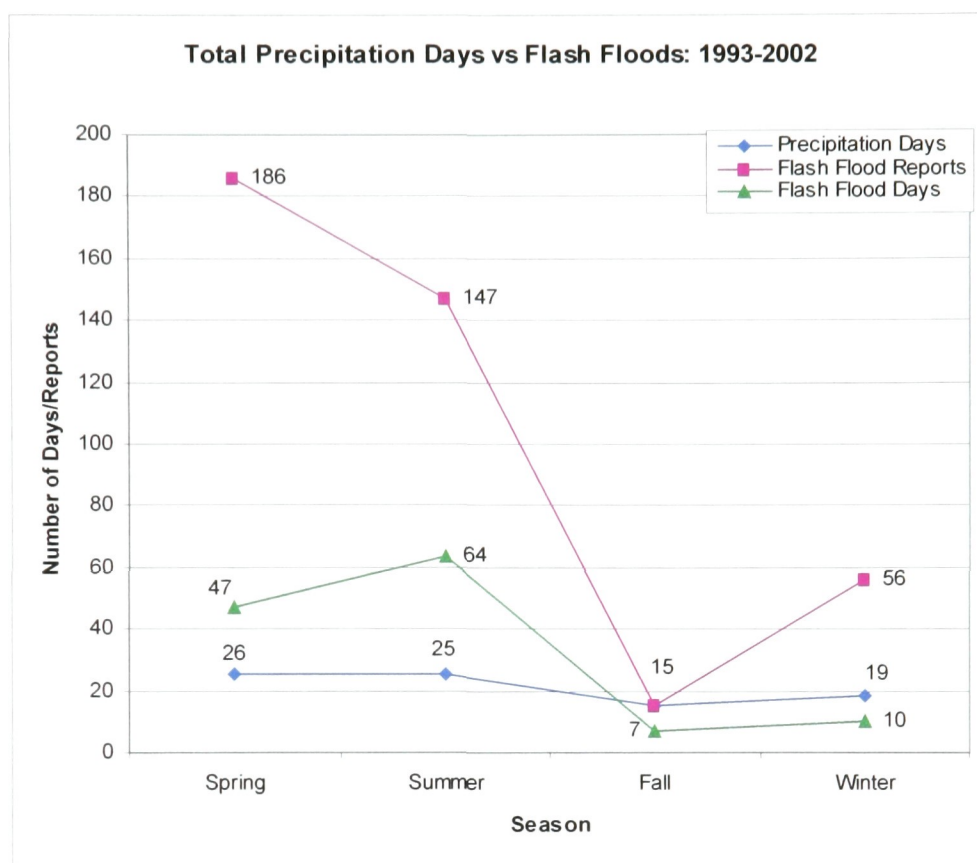




**Figure 6.5: Average monthly precipitation days per station**



**Figure 6.6: Flash Flood Reports by hour**



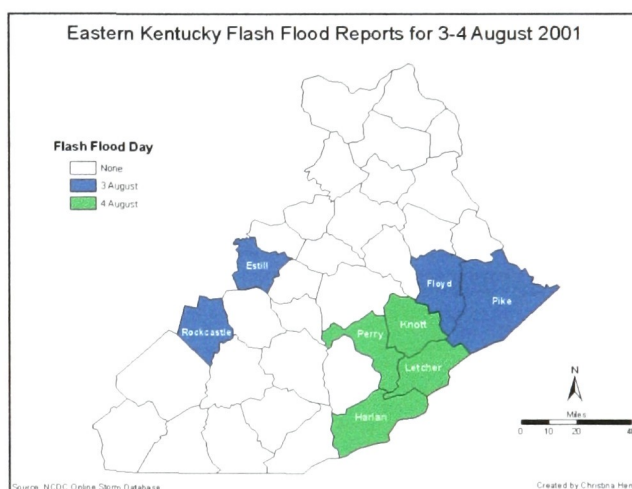
**Figure 6.7: Comparison of precipitation days, flash flood days, and flash flood reports.**

A comparison of flash floods to precipitation days provides some insight as to the nature of flash flooding in Eastern Kentucky (Figure 6.7). During this decade (1993-2002), spring received the most flash flood reports, with 46% of the total reported. The most precipitation days also occurred during the spring, consisting of 31% of the total days. The flash flood days for spring made up only 37% of the total flash flood days, as opposed to the 50% that occurred during the summer. Since spring received the majority of the flash flood reports, the storms producing the necessary rainfall were likely widespread. The slight decrease of precipitation days with an increase in flash flood days suggests that storm systems during the summer moved more slowly than during the Spring, allowing for more flash flooding to occur with less than an 25.4 mm (1 in) of rainfall as the systems left the area. Fall received the fewest flash flood days (5%), flash flood reports (4%), and precipitation days (18%). The storm systems that produced the

flash flooding during the fall were likely smaller scale, causing more localized flash flooding. Other factors that influence the development of flash flooding, and the time of year they occur, include soil moisture, landuse\landcover, terrain, and the influx of moisture.

## VII. EVENT ANALYSIS

A detailed examination of the summertime flash flood event of 3-4 August 2001 analyzes the atmospheric conditions and radar estimates associated with the event. On 3 August 2001, a slow-moving cold front approached Eastern Kentucky triggering heavy rains and flash flooding. Ten reported flash floods affected eight Eastern Kentucky counties (Figure 7.1), causing two injuries and two deaths when the victims were swept into floodwaters. These flash floods resulted in a combined total of \$15.4 million of property damage, rendering Pike, Floyd, Knott, Letcher, and Perry Counties disaster areas (NCDC 2003). Pike and Floyd experienced the worst flash flooding of the event, with a combined total of \$13.6 million in property damage (NCDC 2003). Pikeville reported the highest daily precipitation total for 3 August, at 99 mm (3.9 in; MRCC 2004).

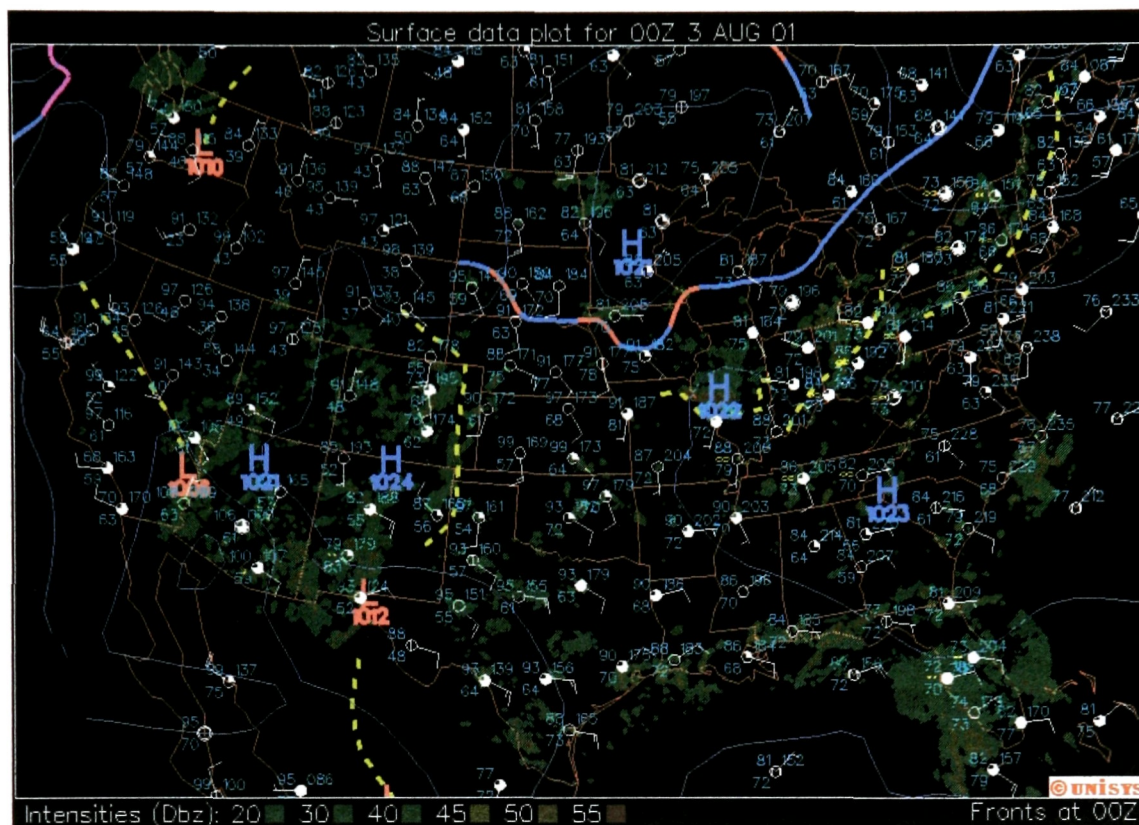


**Figure 7.1: Counties reporting flash flooding**

### 7.1 Large-scale

During the evening of 2 August 2001, a band of clouds and precipitation associated with a slow-moving cold front stretched through Indiana, Ohio, and Northern Pennsylvania, bending northeastward through Maine. Scattered clouds also developed over much of Kentucky, just east of a trough axis (Figure 7.2). The region lay under a large-scale ridge, although there existed



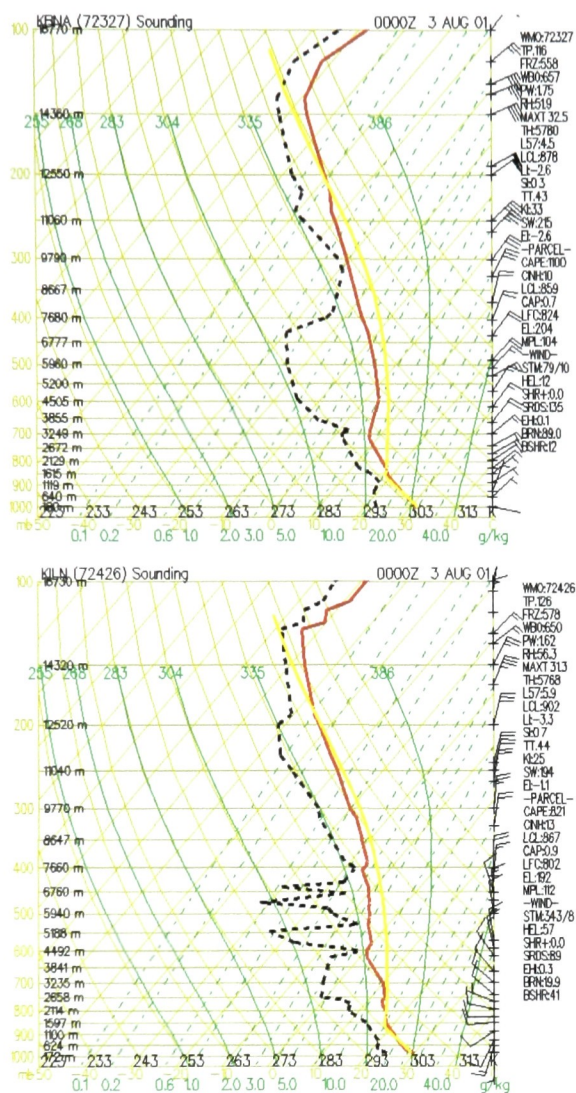


**Figure 7.2: Surface analysis data for 20 EDT 2 August (00Z 3 August; Unisys 2004)**

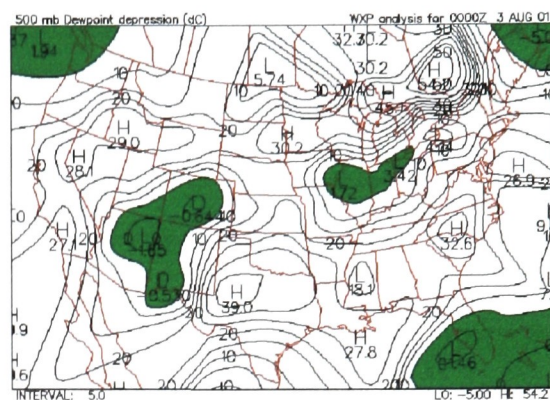
**Table 7.1: Precipitable Water compared to the normal (Climate Diagnostics Center, 2006).**

Station	KBNA	KILN
Regional Average (mm)	33.75	31.25
Percent above Normal		
00Z 3 August 2001	32% (44.45 mm)	32% (41.15 mm)
12Z 3 August 2001	55% (52.32 mm)	67% (52.32 mm)
00Z 4 August 2001	46% (49.28 mm)	39% (43.43 mm)
12Z 4 August 2001	52% (47.50 mm)	-2% (30.48mm)

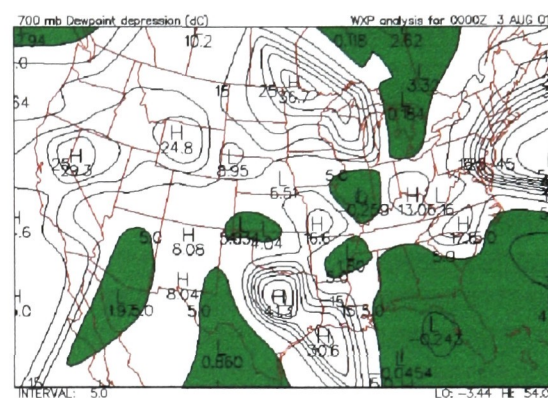
a small region of slightly lower geopotential heights, just west of Kentucky, at 300-mb to 250-mb. The Wilmington, Ohio and Nashville, Tennessee soundings indicated high, lower tropospheric moisture, although both locations recorded a dry middle layer (Figure 7.3). Precipitable water (surface to 500-mb) values for Wilmington (41.15 mm; 1.62 in) and Nashville



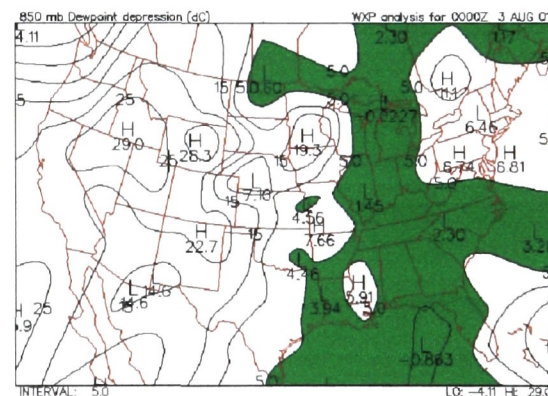
**Figure 7.3: Soundings for Nashville TN (top) and Wilmington OH (bottom) at 00Z 3 August 2001 (PSC 2004)**



a)



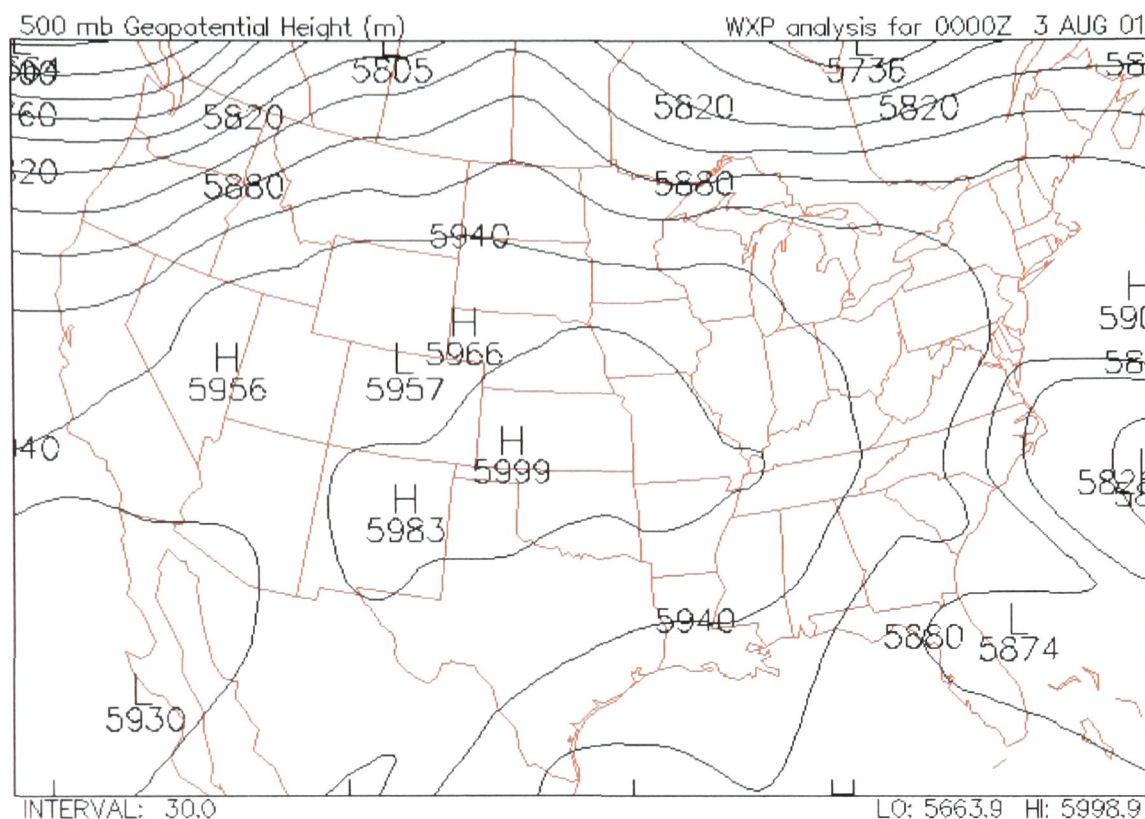
b)



c)

**Figure 7.4: The shaded areas show dewpoint depressions 5°C or less at 00Z, 3 August for a) 500mb, b) 700mb, and c) 850mb (PSC 2004).**





**Figure 7.5: Eastern Kentucky is located on the eastern edge of a large-scale ridge (PSC 2004).**

(44.45 mm; 1.75 in) also indicated a high moisture content, with values 32% above normal (Table 7.1; Figure 7.4).

The surface temperatures in the mid-Atlantic region ranged from the 70's to the 80's °F with dewpoint temperatures primarily in the 60's °F. The winds varied, blowing in from the south and west at approximately 5-10 knots. Weak upper-level winds moved the storm system slowly to the east; Nashville experienced a maximum windspeed of 50 kts (200-mb), while Wilmington experienced a maximum of 35 kts (~160-mb). Meanwhile, a large-scale ridge (Figure 7.5) throughout the troposphere (850-mb – 250-mb) brought winds from the east-northeast into the Nashville area. Wilmington experienced lower-level veering from the south to the north-northeast in the upper-levels.

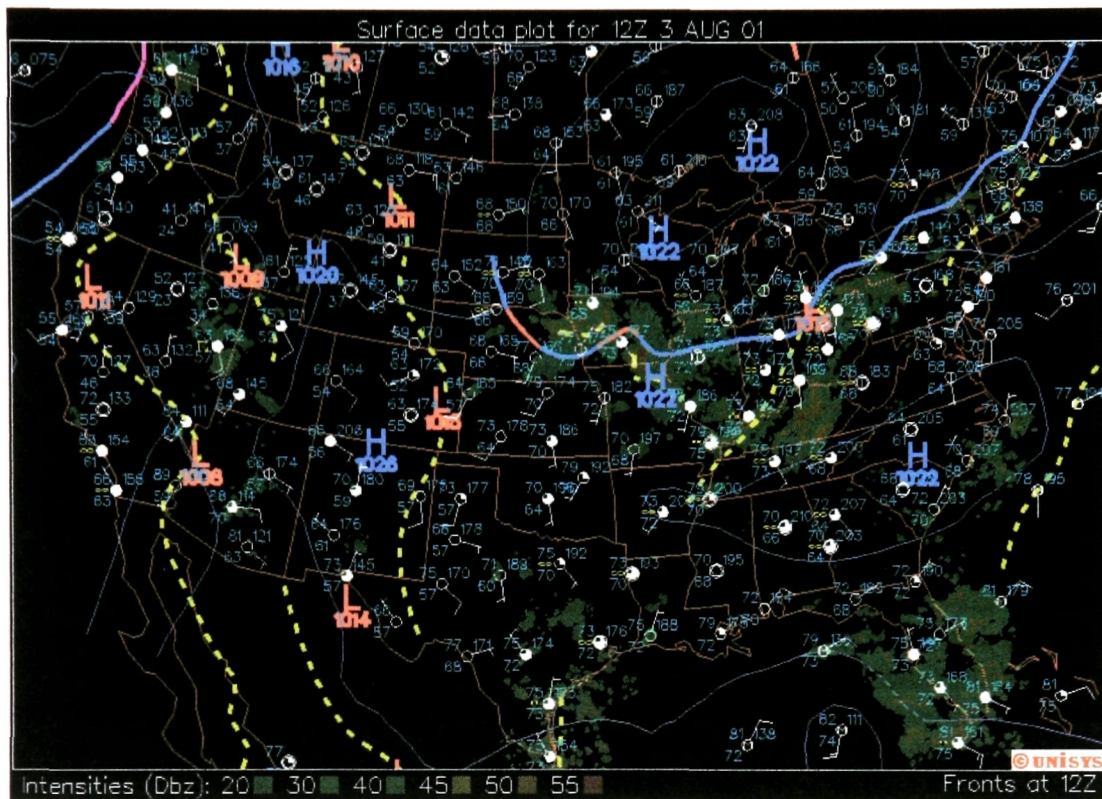
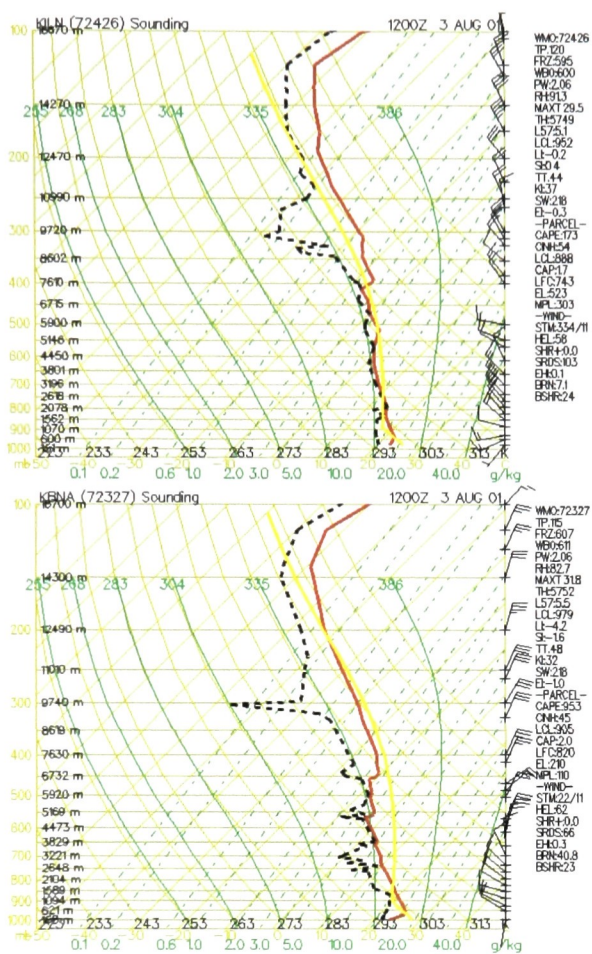


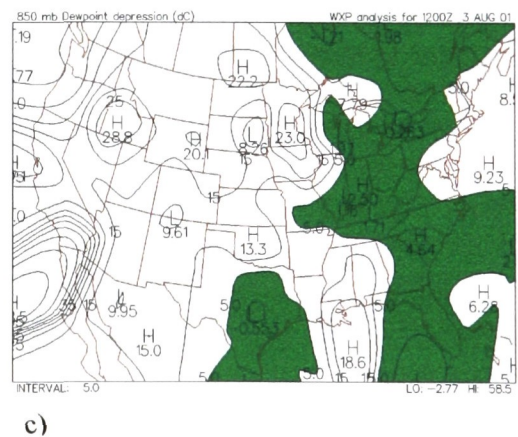
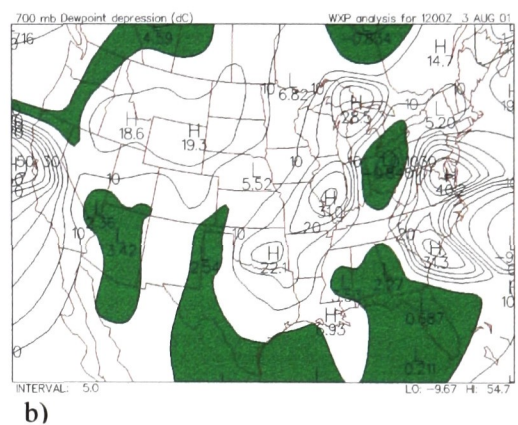
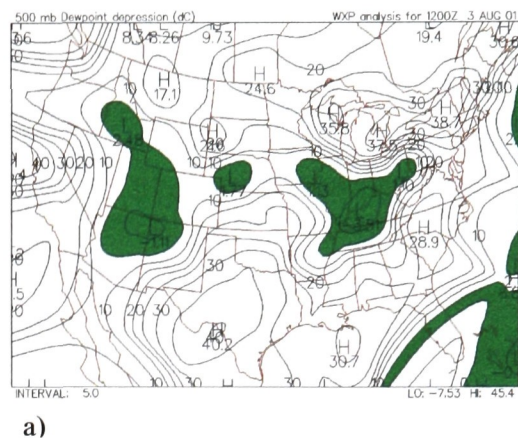
Figure 7.6: Surface analysis data for 08 EDT 3 August (12Z; Unisys 2004)

By early morning on 3 August, the band of precipitation had moved southward, dipping into central Kentucky, and by 0830 EDT (1230Z), precipitation had developed over most of Kentucky (Figure 7.6). Only the farthest eastern portion of Kentucky remained precipitation free. The 0800 EDT (12Z) Nashville and Wilmington soundings indicated an increase in atmospheric moisture, both exhibiting a surface-to-500-mb relative humidity greater than 80% (Figure 7.7 & 7.8). Furthermore, Nashville and Wilmington indicated very high atmospheric moisture with a precipitable water value of 52.32 mm (2.06 in) at both locations, values 55% and 67% (Table 7.1) above normal, respectively. Nashville reported an increase of instability with a lifted index of  $-4.2$ , while Wilmington showed a decrease of atmospheric instability ( $LI = -0.2$ ). However, the K-index indicated instability for both Wilmington (37) and Nashville (32).

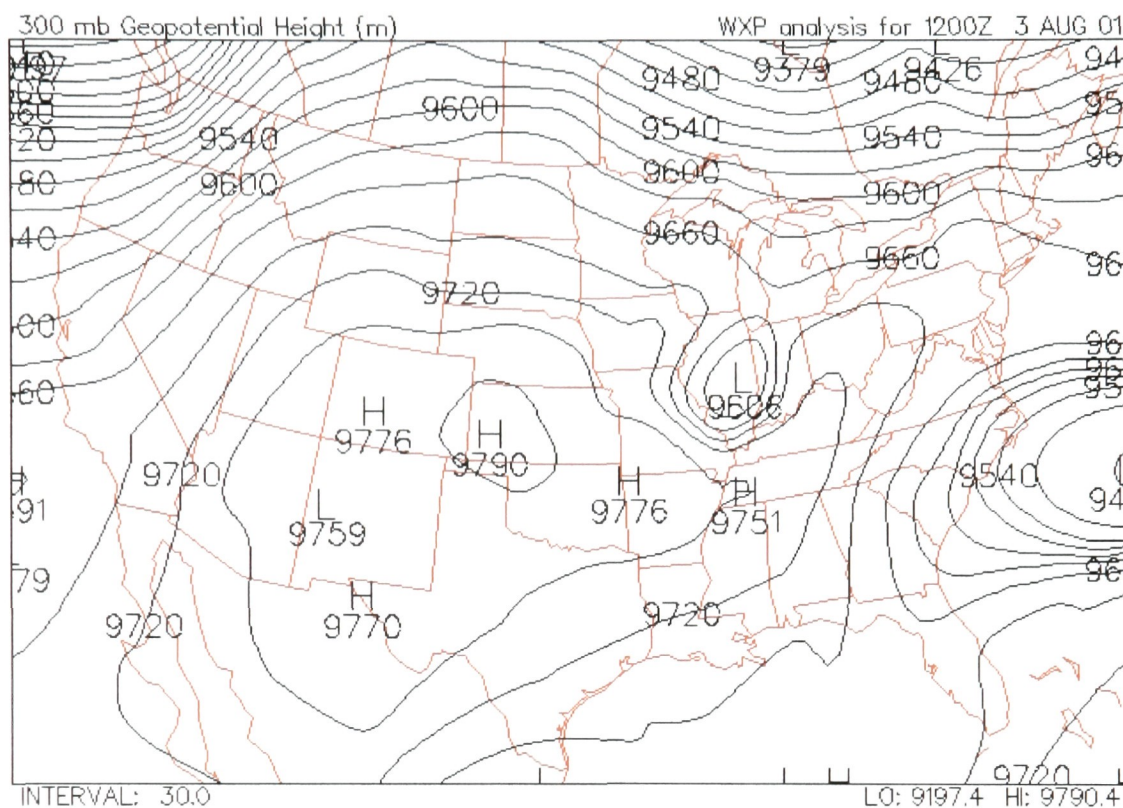




**Figure 7.7: Observed soundings for Wilmington OH (Top) and Nashville TN (bottom) at 12Z 3 August 2001**



**Figure 7.8: The shaded areas show dewpoint depressions 5°C or less at 12Z, 3 August for a) 500mb, b) 700mb, and c) 850mb.**

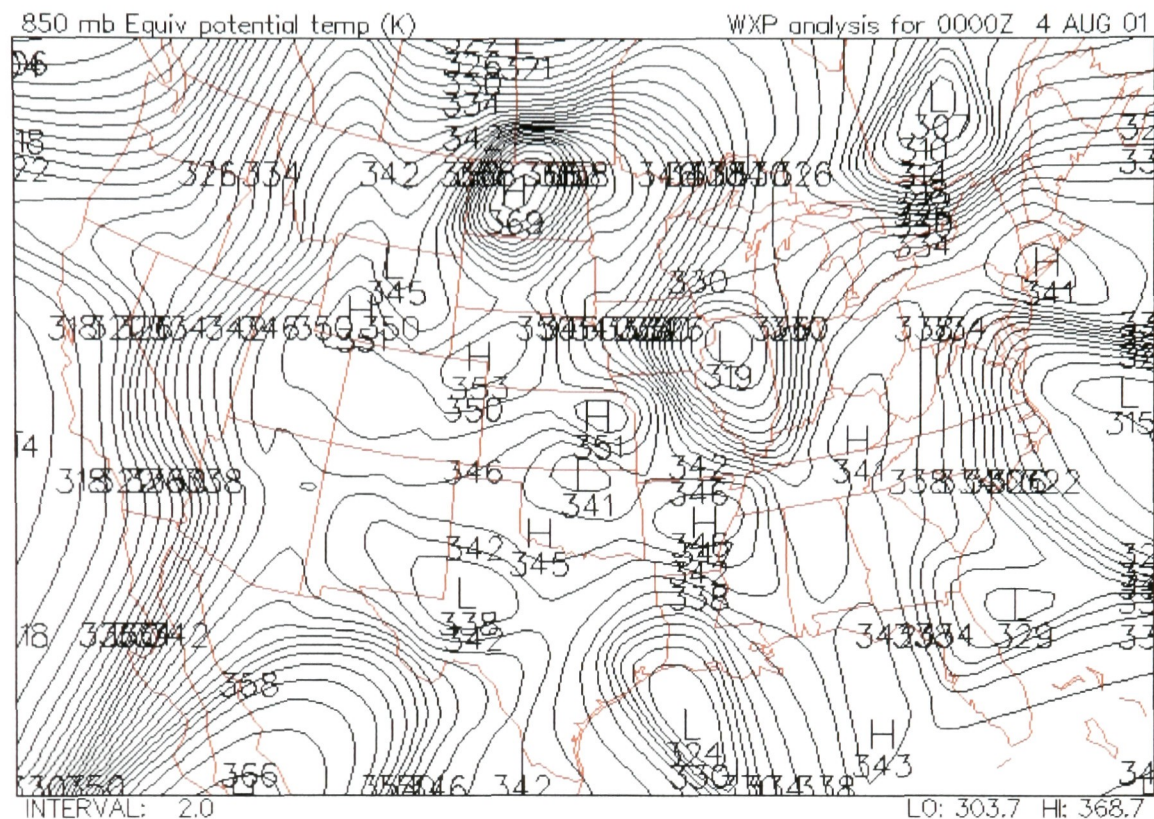


**Figure 7.9: 300-mb Geopotential Heights for 12Z 3 August 2001 (PSC 2004)**

A trough had developed over eastern Michigan, stretching through Indiana and Illinois (Figure 7.9). A large height gradient existed around the trough at 300-mb and 250-mb, which was centered over east-central Illinois. Nashville continued to exhibit vertical directional shear from south to north-northeast, from the surface to about 575-mb. Wilmington, however, experienced west- northwest winds by this time, although the maximum winds for both continued to remain very low (Wilmington = 45 kts; Nashville = 35 kts).

By 1730 EDT (2130Z), the precipitation band had dissipated considerably from Nebraska to Kentucky. However, strong radar intensities existed from northern Mississippi and Alabama, northeastward to southern Maine. Precipitation ended in Illinois, Indiana, and Ohio by 2030 EDT (0030Z 4 August). However, the eastern half of Kentucky had dense cloud cover, which





**Figure 7.10: Equivalent Potential Temperatures for 20 EDT (PSC 2004)**

stretched northeastward to southern Maine, although the precipitation along this cloud band had dissipated somewhat.

The middle layers began drying out by 2000 EDT (00Z 4 August). The lower levels remained moist, however, as indicated by the still high precipitable water values for Nashville (49.28 mm; 1.94 in) and Wilmington (43.43 mm; 1.71 in), values 46% and 38% (Table 7.1) above normal, respectively. The development of a theta-e ridge (Figure 7.10) over the Appalachians of Eastern Kentucky, West Virginia, western Virginia and western North Carolina indicated that the moist air mass had moved eastward ahead of the front, which lay across the northwestern Kentucky border.

Along the Eastern Kentucky-Tennessee border, there existed a vorticity value of  $14.1 \times 10^{-5}$  /s, a value which had more than doubled in the past 12 hours (12Z 3 August – 00Z 4 August).

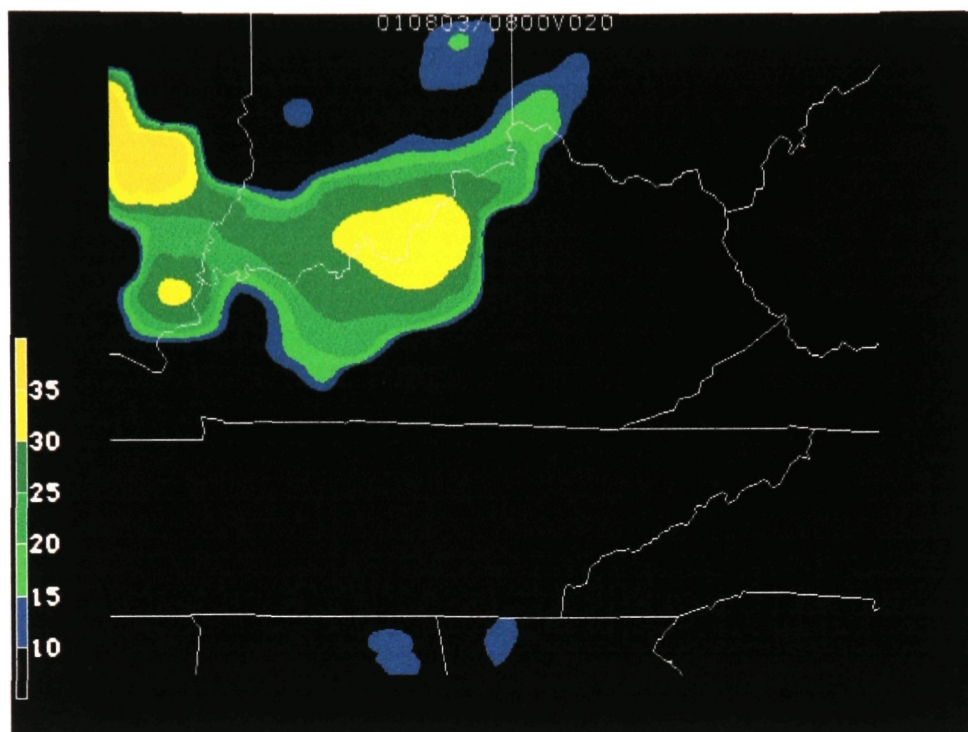
Meanwhile, southwestern Indiana had a vorticity of  $7.73 \times 10^{-5}$  /s. The trough that had existed twelve hours earlier had weakened by 2000 EDT (00Z on the 4<sup>th</sup>). Remnants of the trough remained just north of Kentucky at 850-mb and 700-mb. The atmosphere remained unstable, indicated by the negative lifted indices (Nashville = -2.3; Wilmington = -3.5) and the high K-index values (Nashville = 37; Wilmington = 36). Nashville and Wilmington continued to experience weak west-northwesterly winds, with a maximum windspeed of only 40 kts. The precipitation continued to move slowly eastward with the front, and by 0300 EDT 4 August (0730Z), precipitation ended over the majority of Kentucky.

By 0800 EDT (12Z), the middle- and upper-layers of the troposphere had dried up considerably over Nashville, although the precipitable water value (47.50 mm; 1.87 in) remained high, at 52% above normal (Table 7.1). The lifted index (-2.1) and K-index (38) continued to indicate some atmospheric instability. Wilmington, on the other hand, indicated stable conditions and a very dry atmosphere. The precipitable water had dropped to 30.48 mm (1.2 in), 2% below normal; the lifted index had dropped to -0.0, and the K-index had dropped to 1. Nevertheless, the winds remained weak at 40 kts (Nashville) and 30 kts (Wilmington). Nashville and Wilmington continued to experience northwesterly winds above 450-mb and 350-mb, respectively, with northeasterly winds in the middle and lower-levels. A weak trough dipped south through Eastern Kentucky, at 850-mb; throughout the rest of the troposphere, Eastern Kentucky experienced weak cyclonic circulation.

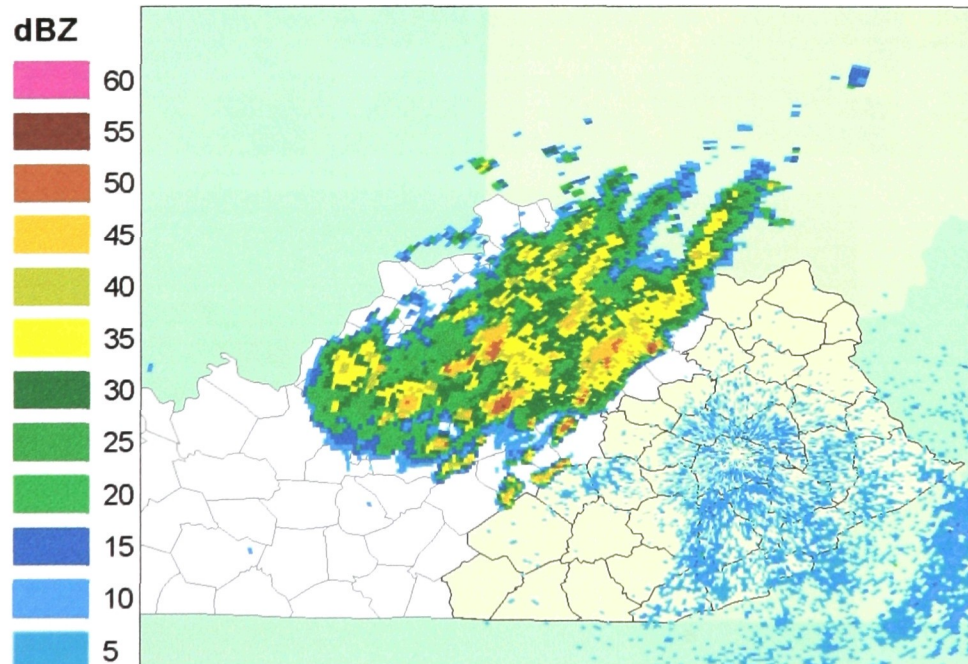
## 7.2 Local-scale

Surface and radar data are the only observed data available for Kentucky. The only upper-air data available for Kentucky comes from Nashville Tennessee and Wilmington Ohio, which does not provide the details needed for a local-scale analysis. Therefore, the upper-air data analyzed in this section comes from the MM5 simulation discussed earlier (Chapter 5.3b).





**0438 EDT (0838 Z) - 3 August 2001**



**Figure 7.11: Modeled Reflectivity for 0400 EDT (Top) and Observed radar reflectivity at 0438 EDT (bottom) indicating rain moving into Eastern Kentucky.**

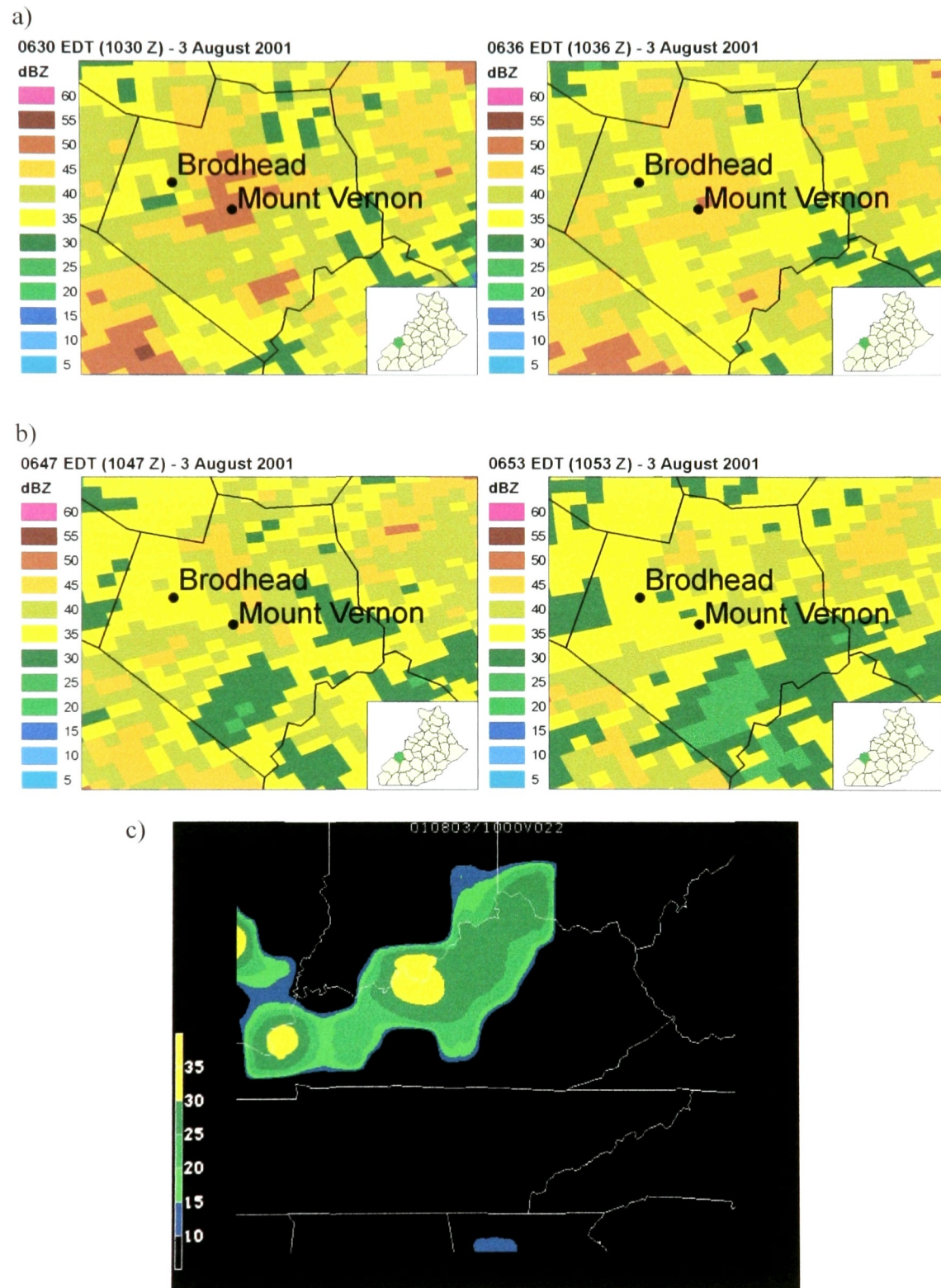
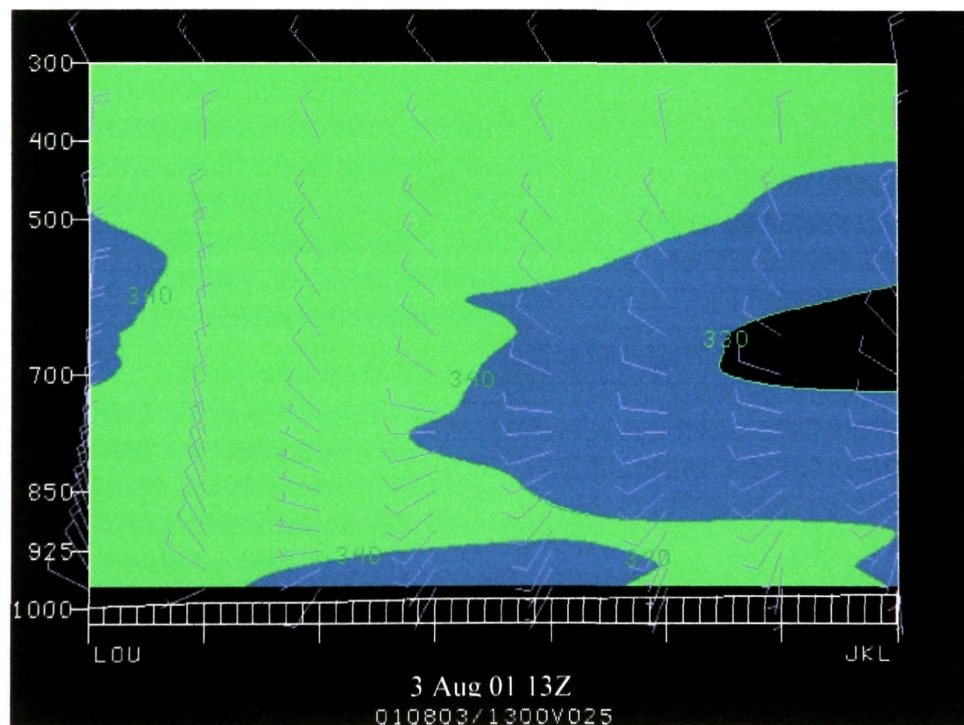
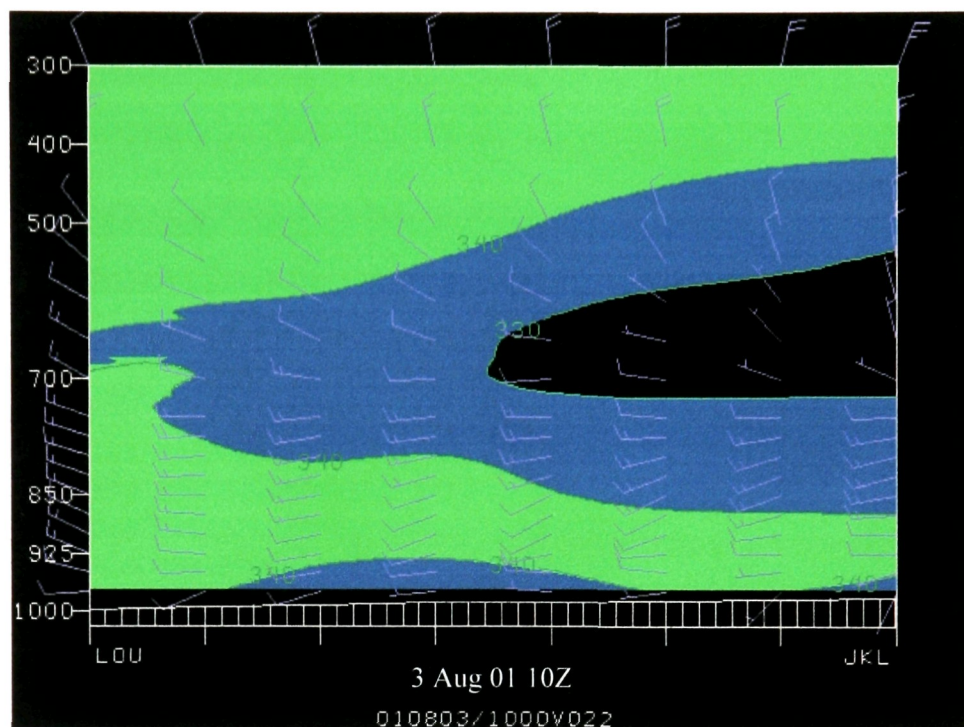


Figure 7.12: Jackson KY Level III Doppler Radar images for Rockcastle County at a) 0630-0636 EDT and b) 0647-0653 EDT, c) Modeled radar reflectivity at 0600 EDT.





**Figure 7.13: Modeled wind barbs and Theta-e cross-section from Louisville to Jackson (0600-0900 EDT 3 August 2001)**

Moderate to heavy rain fell across central Ohio during the evening of 2 August 2001. However, stations throughout Eastern Kentucky reported clear skies and calm, hazy conditions. Temperatures in Jackson and London Kentucky fell from the lower 80s °F to the mid- to lower 70s °F. Consequently, the dewpoint depressions decreased from the mid- to upper teens, to only a few degrees Fahrenheit. The surface pressure remained relatively constant at approximately 1021-mb.

The surface conditions changed little during the early morning hours of 3 August. However, a line of showers moved slowly south-southeastward toward Eastern Kentucky from Indiana. At 0427 EDT (0827Z), the radar indicated the development of showers at the border of Rockcastle County, ahead of the existing line (Figure 7.11). These showers developed rapidly, and by 0635 EDT (1035Z), the heavy rains led to two reports of flash floods in Rockcastle County, as water began to cover Highways 25 and 150, near Conway, and numerous roads in Mt Vernon (Figure 7.12a). In response to the reports, the National Weather Service in Jackson issued a flash flood warning for Rockcastle County at 0636 EDT (1036Z). Rain continued to fall, leading to a third flash flood report in Rockcastle County at 0650 EDT (1050Z), where Highways 70 and 150, near the town of Brodhead, were closed due to high water (Figure 7.12b).

The atmospheric moisture over Eastern Kentucky continued to increase, and by 0800 EDT (12Z), the dewpoint depressions for both Jackson and London had dropped to 0°F, with temperatures and dewpoint temperatures at 70°F. Above normal precipitable water values for Pikeville (37 mm) and Prestonsburg (38 mm) and rising theta-e values (Figure 7.13) for Jackson (330-340 K) also indicated increasing moisture. By 0900 EDT (13Z), London and Jackson reported light rain. In fact, light to moderate rain fell over much of southwestern Eastern Kentucky, and at 0934 EDT (1334Z), the National Weather Service issued a flash flood watch for the area. At 0943 EDT (1343Z), the flash flood watch was amended to include Harlan and Clay Counties (Figure 7.14).



**Figure 7.14: Flash Flood Watch**

Atmospheric moisture continued to rise throughout the morning and into the afternoon. The temperature at Jackson rose slightly to a maximum of 72°F, while the temperature at London rose to a height of 74°F by early afternoon. The dewpoint temperatures fluctuated slightly, though remaining in the lower 70s °F. The dewpoint depressions for both stations remained low, under 5°F. The precipitable water for Jackson (48 mm) and Pikeville (38 mm) also continued to rise with values 54% and 23% above normal, respectively (Table 7.2).

By 1500 EDT (19Z), the precipitable water had reached 49 mm at Jackson, and 40 mm at Prestonsburg and Pikeville – values 58% and 29% above normal, respectively. The temperatures and dewpoint temperatures remained in the low 70's °F with dewpoint depressions of 2°F at Jackson and 3°F at London. Jackson reported light rain, while London reported haze. Doppler radar indicated only scattered showers and at 1537 EDT (1937Z), based

**Table 7.2: Precipitable Water for Eastern Kentucky compared to the regional average (Climate Diagnostics Center, 2006).**

Station	Jackson	Prestonsburg	Pikeville
Regional Average (31.25 mm)	Percent above Normal		
11 EDT (15Z)	37% (43 mm)	15% (36 mm)	16% (36 mm)
12 EDT (16Z)	42% (44 mm)	19% (37 mm)	19% (37 mm)
13 EDT (17Z)	48% (46 mm)	20% (38 mm)	22% (38 mm)
14 EDT (18Z)	54% (48 mm)	NA	23% (38 mm)
15 EDT (19Z)	58% (49 mm)	29% (40 mm)	28% (40 mm)
16 EDT (20Z)	60% (50 mm)	37% (43 mm)	34% (42 mm)
17 EDT (21Z)	60% (50 mm)	45% (45 mm)	43% (45 mm)
18 EDT (22Z)	56% (49 mm)	48% (46 mm)	47% (46 mm)
19 EDT (23Z)	51% (47 mm)	46% (46 mm)	46% (46 mm)
20 EDT (00Z)	49% (47 mm)	44% (45 mm)	45% (45 mm)

on analysis of conditions, the National Weather Service canceled the flash flood watch for southwestern Eastern Kentucky. Though there remained a chance for heavy rainfall, major problems were not anticipated.

Meanwhile, the theta-e continued to increase over Jackson and by 1700 EDT (21Z), the theta-e had reached its highest level. The local and regional theta-e values ranged from less than 350 K at 850-mb, less than 340 K at 700-mb, and 340-350 K at 500-mb. The precipitable water continued to increase, reaching 50 mm (2 in) at Jackson and 45 mm at both Prestonsburg and Pikeville – values 60% and 44% above normal, respectively (Table 7.2; Figure 7.15). Moderate rain fell over portions of Pike and Floyd Counties. By 1800 EDT (22Z) the atmospheric moisture began to decrease over Jackson, as seen in the falling although the precipitable water remained very high at 56% above normal. The surface moisture remained constant at Jackson and London, while the atmospheric moisture continued to rise over Pikeville and Prestonsburg, with precipitable water values increasing to their maximum, at 46 mm – 47% above normal (Table 7.2).

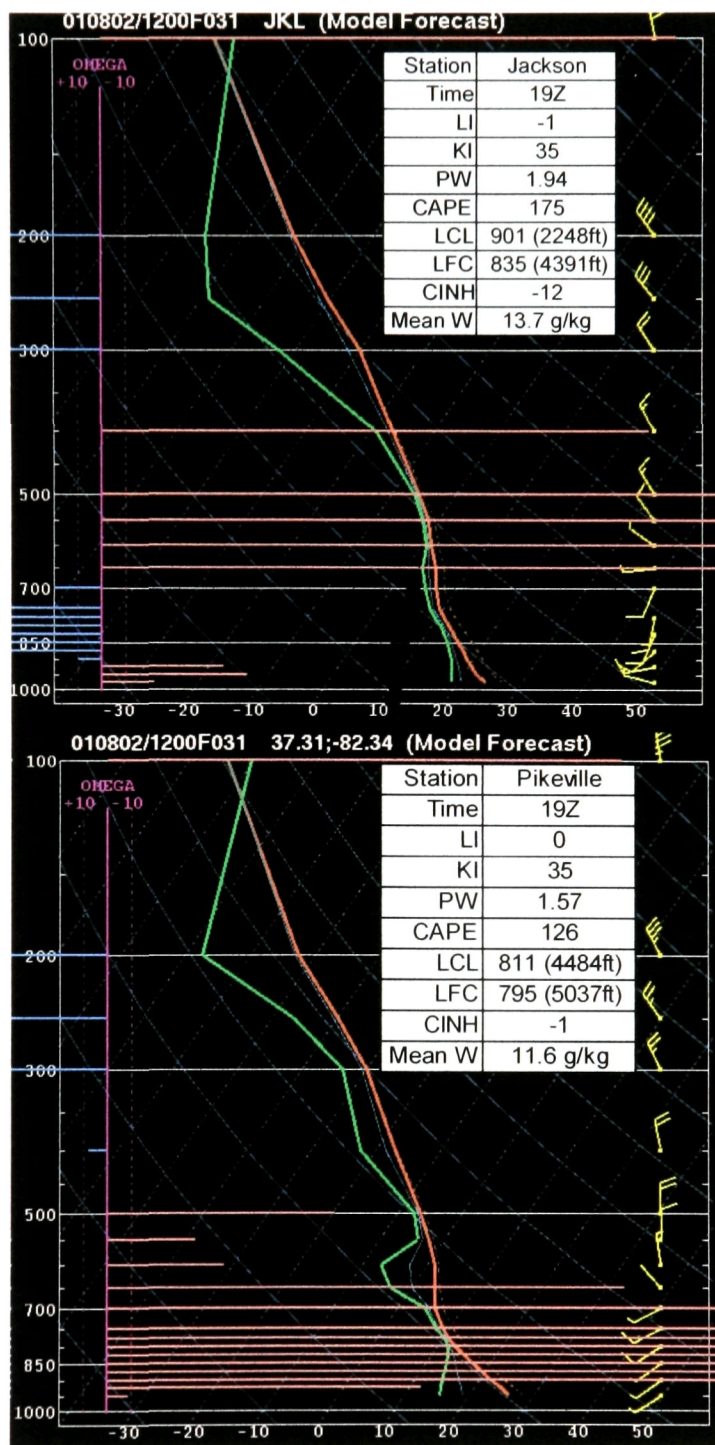


Figure 7.15: Modeled Sounding for Jackson KY (top) and Pikeville KY (bottom) at 15 EDT (19Z) 3 August 2001



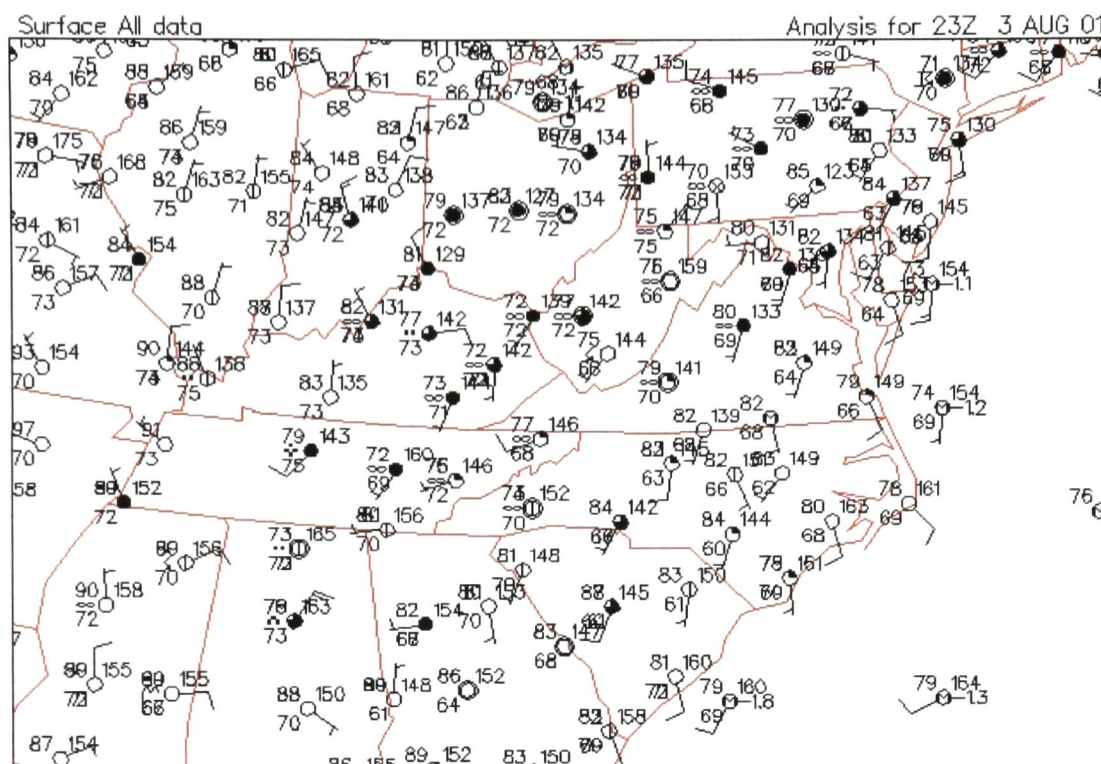


Figure 7.16: Observed surface analysis for 15 EDT (19Z) 3 August 2001 (PSC, 2004).

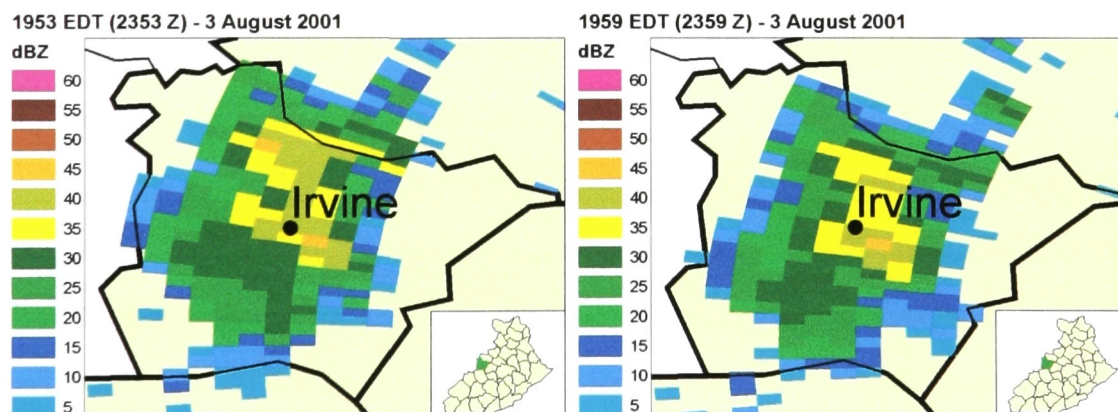


Figure 7.17: Jackson KY Level III Doppler Radar images for Estill County at 1953-1959 EDT



By 1900 EDT (23Z), the atmospheric moisture began to slowly decrease over Prestonsburg and Pikeville, although both atmospheric and surface moisture remained high. The temperatures and dewpoint temperatures remained in the low 70's °F, giving Jackson and London dewpoint depressions of 0°F and 2°F, respectively (Figure 7.16). Moderate to heavy rains began falling across central Estill County, and continued for about an hour. By 2000 EDT 3 August (00Z 4 August), two feet of water covered Cow Creek Road approximately four miles east of Irvine (Figure 7.17), leading to another flash flood report.

By 2110 EDT (0110Z), heavy rain fell between Justell and Fishtrap Lake, moving southward at five miles per hour. Five minutes later, flash flooding was reported in Floyd County (Figure 7.18a). At 2116 EDT (0116Z), the National Weather Service issued a flash flood warning for Pike and the eastern portion of Floyd Counties. At 2130 EDT (0130Z), a flash flood report stated that high water had pushed a vehicle off State Route 1384 near Hurricane Creek, just northwest of Pikeville (Figure 7.18b). At 2200 EDT (02Z), Hurricane Creek swept another vehicle off a road near Boldman, resulting in one fatality. At 2230 EDT (0230Z), moderate to heavy rains continued to fall in southern Floyd and Pike counties, with approximately 110 mm (4 in) of rain reported by an observer near Coal Run.

At 23 EDT (03Z), temperatures and dewpoint temperatures remained in the low 70's°F, with dewpoint depressions for Jackson and London at 0°F and 1°F, respectively. By 2319 EDT (0319Z), heavy rain fell over eastern Knott County. The storms moved slowly south at 10 mph, and at 2330 EDT (0330Z), the National Weather Service issued a flash flood warning for eastern Knott County (Figure 7.19).

Two hours later, at 0130 EDT, 4 August (0530Z), high water was reported on Highway 160, near Brinkley. At 0158 EDT (0558Z), the National Weather Service issued a flash flood

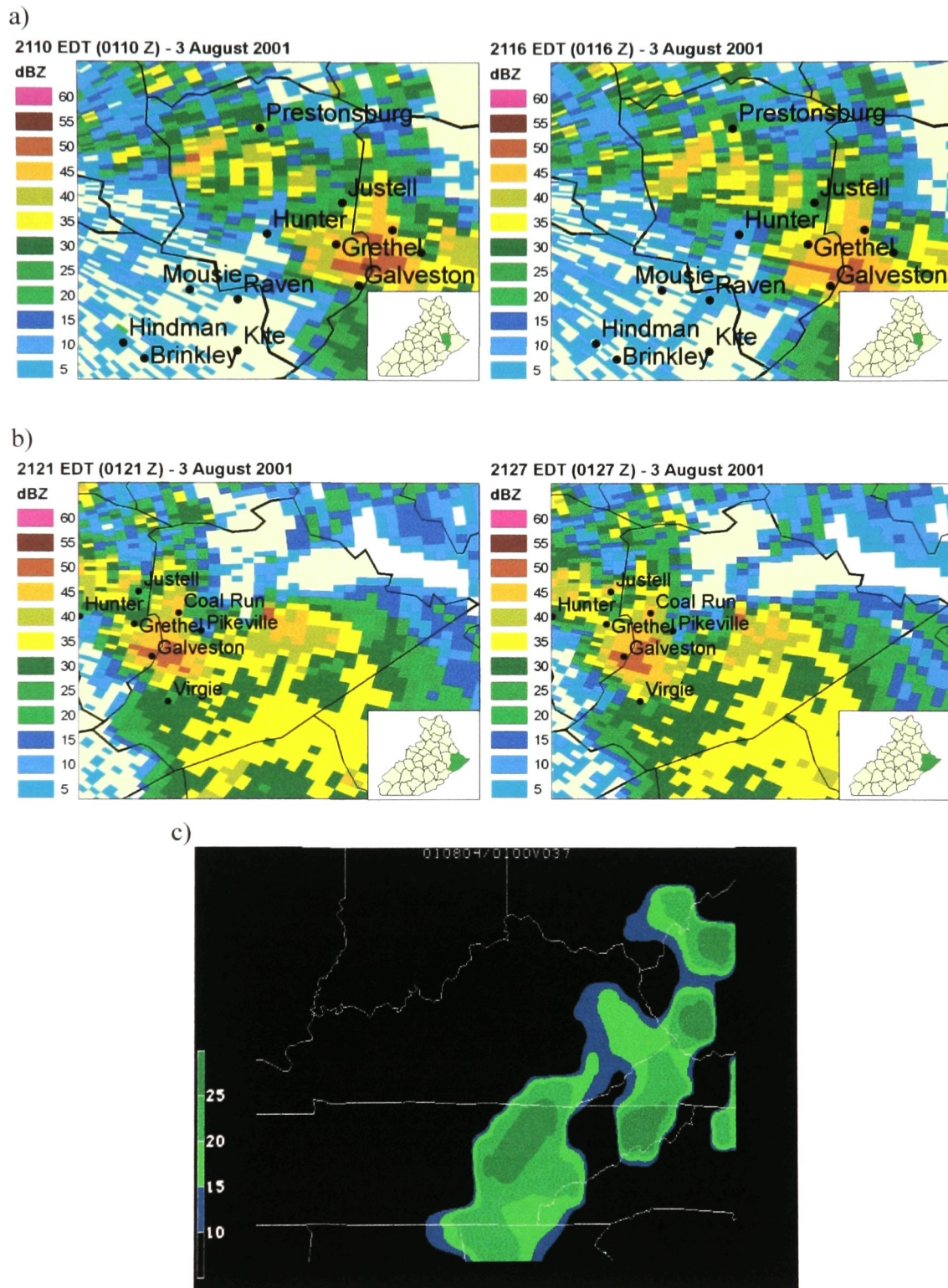


Figure 7.18: Jackson KY Level III Doppler Radar images for a) Floyd County at 2110-2116 EDT and b) Pike County at 2121-2127 EDT, and c) Modeled radar reflectivity at 2100 EDT.

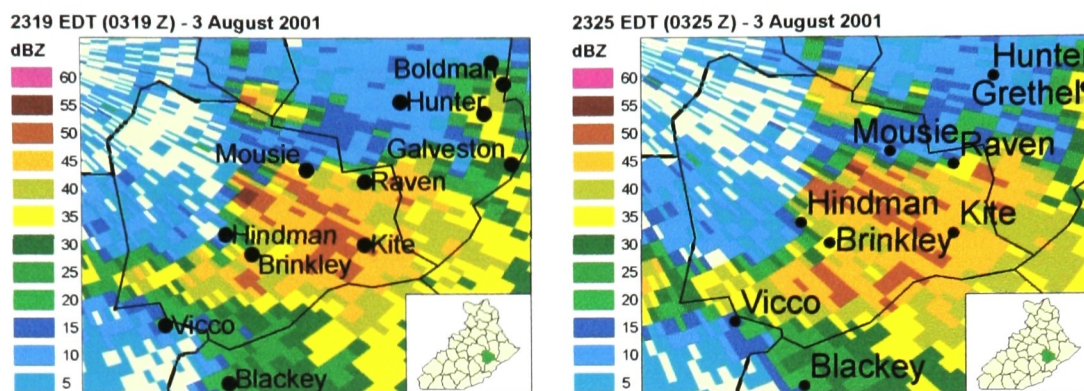


Figure 7.19: Jackson KY Level III Doppler Radar images for Knott County at 2319-2325 EDT

warning for Letcher and southern Perry Counties. At 0200EDT (06Z), both Jackson and London had dewpoint depressions of 0°F. Doppler radar indicated moderate rainfall in portions of Perry, Knott, Letcher, and Harlan Counties. By 0215 EDT (0615Z), the National Weather Service reported that areas near Hindman, Brinkley, and Redfox had already received 76-102 mm (3-4 in) of rain.

At 0230 EDT (0630Z), flash flooding was reported in Perry and Knott Counties as moderate to heavy rains fell. In Perry County, flooding was reported in Vicco where two families were trapped by high water. In Knott County, five feet of water covered Highway 550, at Hindman. Consequently, the flash flood warning for Knott County was extended. At 0235 EDT (0635Z), Highway 7 near Blackley (Letcher County) experienced flooding, and at 0245 EDT (0645Z), flooding was reported along Elk Creek, at Road 3408 and Highway 7.

By 0300 EDT (07Z), only scattered showers continued in portions of Pike and Floyd Counties. Therefore, the flash flood warnings were allowed to expire. However, Doppler radar indicated moderate rains in Harlan County, particularly near Totz, Leonard, and Shields. At 0308 EDT (0708Z), the National Weather Service issued a flash flood warning for Harlan County. By 0330 EDT (0730Z), Perry and Letcher counties continued to experience scattered showers, which were expected to end within a few hours. By 0500 EDT (09Z), the rain had ended in Knott

county and the flash flood warning expired. However, flooding continued along Highway 510, near Gilley and Gordon, in Letcher County. The dewpoint depressions remained 0°F at Jackson and London. By 0652 EDT (1052Z), the rains had ended in Perry, Harlan, and Letcher counties, and the flash flood warnings for all three counties expired at 0700 EDT (11Z).

## VIII. CONCLUSION

Limited research exists concerning flash flooding of the Appalachian region and Eastern Kentucky. Studies have shown that flash floods generally occur under relatively benign conditions, which makes forecasting them a challenge. Close examination of a wide variety of flash flood events throughout the Contiguous United States indicates that they typically accompany slow-moving systems and occur in regions of high atmospheric moisture. This was also the case with the 3-4 August 2001 Eastern Kentucky flash flood event.

Flash flood warning records for this event were only available for nine of the ten flash flood reports. From the data available, the only flash flood watch issued by the National Weather Service covered the southwestern counties of the study area during the morning of 3 August 2001. Only two of these counties received flash flooding; Rockcastle County received flash flooding prior to the issuance of the watch and Harlan County received flash flooding during the morning of the 4<sup>th</sup>. Of the nine flash flood warnings, the National Weather Service only issued five warnings prior to actual flooding. The other four warnings were not issued until after flash flooding was reported.

In Pike County, where the two fatalities occurred, a flash flood warning was issued 14 minutes prior to the flooding. A warning was issued 32 minutes prior to the flash flooding in Perry County. However, one injury occurred when a man was swept away through two culverts. Though the warning was not issued until one minute after the flash flood report in Floyd County, an injury occurred 44 minutes after the warning to a man whose truck was swept into a river. In each of these cases, warnings had been issued prior to the casualties, despite the relatively benign conditions.



Only marginal instability accompanied the Eastern Kentucky flash flood event. This was shown in the lifted indices, which were  $-1$  throughout the event at Jackson, Kentucky. The Fort Collins, Colorado and Madison County, Virginia storms also had marginal instability with lifted indices of  $-2.8$  and  $-1$ , respectively. Maddox *et al.* (1979) and Gaffin and Hotz (2000) all reported negative lifted indices as flash flood conducive. These conditions were also found during the Big Thompson storm and the Madison County storm, as well as others.

Although some instability existed during the Eastern Kentucky flash flood event, severe storms as described by Miller (1972) were not expected. In fact, all the parameters examined indicated a weak chance for severe weather, except for the dewpoint temperature, which indicated a strong chance for severe weather (Table 8.1). The model data provided the same results.

**Table 8.1: Severe weather characteristics indicate a weak chance for severe weather. Dewpoint temperatures, however, indicate a strong chance for severe weather. The pressure drops indicated a slightly moderate chance for severe weather (Miller 1972).**

	00Z 3 August 2001	12Z 3 August 2001	00Z 4 August 2001	12Z 4 August 2001
	KBNA			
Stability - Totals Index	43.3	47.5	45.7	44.5
500-mb Wind Speed (kt)	17	17	24	17
850-mb Wind Speed (kt)	7	11	15	10
850-mb Dewpoint (°C)	<b>15</b>	<b>15</b>	<b>15</b>	<b>15</b>
12-h Surface Pressure Drop (mb)		2	5	1
Surface Pressure over Threat Area (mb)	1021	1019-1020	1014	1013-1014
Surface Dewpoint (°F)	<b>75.2</b>	<b>71</b>	<b>75.2</b>	<b>71.6</b>
	KILN			
Stability - Totals Index	43.9	43.7	45.1	44.6
500-mb Wind Speed (kt)	12	16	15	18
850-mb Wind Speed (kt)	9	14	18	16
850-mb Dewpoint (°C)	<b>13.7</b>	<b>14.5</b>	<b>13.7</b>	<b>13.2</b>
12-h Surface Pressure Drop (mb)		2	5	1
Surface Pressure over Threat Area (mb)	1021	1019-1020	1014	1013-1014
Surface Dewpoint (°F)	<b>74.8</b>	<b>68.9</b>	<b>71</b>	<b>66.38</b>



Maddox *et al.* (1979) identified a slow-moving system associated with each type of flash flood examined. In Rapid City 1972, a slow-moving front broke a low-level cap over the Black Hills, which caused flash flooding. A mesoscale convective system caused flash flooding in Texas in May 1981. Similarly, the Eastern Kentucky flash floods of August 2001 accompanied a slow-moving cold front, with heavy rains occurring ahead of a trough axis.

Areas of low dewpoint depressions, high precipitable water, and high theta-e values indicate areas prone to heavy rain and flash flooding. Eastern Kentucky had surface dewpoint depressions of only a degree or two throughout the event. Dewpoint depressions were generally 5°C or lower throughout the troposphere. Based on Miller's (1972) severe weather chart, the low dewpoint depressions experienced during the Eastern Kentucky flash flood event indicated a strong chance for severe weather. Madison County, Virginia also experienced low dewpoint depressions. Not only did Eastern Kentucky have low dewpoint depressions; it also had high theta-e values. In fact, a theta-e ridge existed just west of the Appalachians. Furthermore, Eastern Kentucky had above normal precipitable water values, also an indication of very moist air. Maddox *et al.* (1979) noted in their climatology that above normal precipitable water values is a common characteristic associated with flash flooding. The Big Thompson and the Madison County storms, among others, reported above normal precipitable water values.

The Eastern Kentucky flash flood event does not particularly match any of the climatological patterns described in chapter 2. Of these patterns, it most closely resembles the Frontal pattern described by Maddox *et al.* (1979). This type of pattern typically has a west-east oriented frontal boundary. The Eastern Kentucky event did have a west-to-east oriented frontal boundary during 3 August. However, as the front moved eastward, it stalled along the Appalachians and pivoted, becoming southwest-to-northeast oriented. Nevertheless, the front was slow-moving, becoming quasi-stationary as it approached the Appalachians. This event, as with most Frontal events, possessed high atmospheric moisture and weak winds.

The Eastern Kentucky event also exhibited characteristics common to the Pattern I and II heavy rain events described by Konrad (1997), including above normal precipitable water, high atmospheric moisture, K-index values greater than 30, and negative lifted index values. Gaffin and Hotz (2000) also identified these characteristics as common ingredients for flash flooding in the southern Appalachians of North Carolina and Tennessee.

As seen from this research, the atmospheric conditions associated with flash flooding vary depending on season and geographic location. Flash flood climatologies identify the atmospheric patterns associated with flash flooding, providing a blueprint for identifying the potential for flash flooding. Although some climatologies currently exist, a climatology specific to Eastern Kentucky would aid local forecasters in the identification of flash flood conditions and assist in the timely issuance of watches and warnings.

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